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Sr Isotope Composition in Carbonates of the Karatau Group, Southern Urals, and Standard Curve of $^{87}Sr/^{86}Sr$ Variations in the Late Riphean Ocean

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Abstract—The presented Rb–Sr systematics of carbonates from the Karatau Group, the Upper Riphean stratotype of southern Urals, elucidates important details of secular variations of $^{87}Sr/^{86}Sr$ ratios in the Late Riphean seawater, which have been formerly unknown. Samples selected for analysis satisfy all the strict geochemical criteria characterizing the least altered carbonate rocks (Mn/Sr ≤ 0.2, Fe/Sr ≤ 5.0, Mg/Ca ≤ 0.024 in limestones and Mn/Sr ≤ 1.2, Fe/Sr ≤ 3.0, Mg/Ca ≥ 0.608 in dolostones), and they all have been preliminary treated in 1N solution of ammonium acetate for a partial removal of epigenetic carbonate phases. A verified curve of secular variations of $^{87}Sr/^{86}Sr$ ratio in the Late Riphean ocean is plotted based on new data and Sr isotope parameters formerly known for carbonates of the Upper Riphean key sections. As is established, that ratio was nearly constant, ranging from 0.70519 to 0.70566, within the time span of 1030–810 Ma and next rose up to 0.70611 about 775 Ma ago. Afterward, between 765 and 740 Ma, it decreased down to 0.70561–0.70575 and then, within the time span of 740–690 Ma, it ranged from 0.70646 to 0.70686 with a short-term drop down to 0.70620 about 720 Ma ago. At the end of the Late Riphean (660–640 Ma), the ratio lowered to 0.70538–0.70580 to become growing up to 0.70840–0.70860 during the Vendian and initial Cambrian. The established variations of the $^{87}Sr/^{86}Sr$ ratio in the Late Riphean ocean have been controlled by a combination of geodynamic factors, magmatic events, sea-level oscillations, and compositional changes in provenances, which were subjected to erosion at that time. An additional influence of climatic fluctuations over supercontinent and its fragments is also admissible.

Key words: $^{87}Sr/^{86}Sr$ ratio in seawater, Late Riphean, carbonate rocks; Karatau, Atar, Shaler, and Akademikerbreen groups, Bitter Springs Formation.

INTRODUCTION

Investigation of secular variations of $^{87}Sr/^{86}Sr$ ratio in paleooceans is an important tool for reconstruction of past geodynamic environments. In addition, this approach elucidates the crustal erosion intensity and terrigenous influx composition at various stages of the Earth evolution, and it may be used as well in correlation of carbonate successions, when the ratio gradients are high. The following aspects define effectiveness of such investigations. (1) At the each moment of geological history, the $^{87}Sr/^{86}Sr$ ratio is constant in the World Ocean, because the Sr residence time in seawater is three orders of magnitude greater than the stirring time of the whole water mass (Goldberg, 1963; Faure, 1986; Aberg and Wickman, 1987; Hodell *et al.*, 1989). (2) There are three variables that control secular variations of the ratio: the Sr flux with low $^{87}Sr/^{86}Sr$ ratios from the mantle, the continental runoff intensity, and the Sr isotope composition in the latter (Faure *et al.*, 1965; Veizer and Compston, 1974; Palmer and Edmond, 1989; Richter *et al.*, 1992; Jacobsen and Kaufman, 1999; Semikhatov *et al.*, 2002). (3) Ca-bearing minerals of marine origin (carbonates fist of all) inherit the Sr isotope composition of sedimentation medium, and their primary phases are able to retain this isotopic signature until the recent time (Peterman *et al.*, 1970; Tremba *et al.*, 1975; Koepnick *et al.*, 1985).

A detailed and internally consistent curve of seawater $^{87}Sr/^{86}Sr$ variations has been plotted for the Phanerozoic evolution of the Earth (Burke *et al.*, 1982; Koepnick *et al.*, 1985; Denison *et al.*, 1994; Veizer *et al.*, 1999; McArthur *et al.*, 2001, and references therein). Considerable intervals of this curve are suitable for stratigraphic correlation and even for high-precision chronological calibration of carbonate successions (DePaolo and Ingram, 1985; McArthur, 1994; Diener *et al.*, 1996; Ebneth *et al.*, 2001; Melezhik *et al.*, 2002). Some extremes of the curve are also found to be corresponding to concrete geodynamic events in the past

(Hodell *et al.*, 1990; Richter *et al.*, 1992; Godderis, Francois, 1995).

Investigation of seawater $^{87}Sr/^{86}Sr$ variations in the Proterozoic ocean that had been initiated later (Veizer, Compston, 1976; Veizer *et al.*, 1983) became active in the last 10–12 years only (Derry *et al.*, 1989, 1992; Asmerom *et al.*, 1991; Pokrovskii and Vinogradov, 1991; Kaufman *et al.*, 1993, 1997; Mirota and Veizer, 1994; Gorokhov *et al.*, 1995, 1998; Gorokhov *et al.*, 1996; Kuznetsov *et al.*, 1997; Semikhatov *et al.*, 1998, 2002; Vinogradov *et al.*, 1998; Walter *et al.*, 2000; Bartley *et al.*, 2001, and references therein). Nevertheless, even the seawater $^{87}Sr/^{86}Sr$ curve plotted for the Late Riphean interval that is studied better than others is not consistent so far (Melezhik *et al.*, 2001). This is a consequence of the following objective and subjective issues: there are known few well-dated carbonate successions of the Proterozoic; significance of epigenetic alterations in carbonate rocks is underestimated; criteria used to control the preservation state of Rb–Sr isotopic systems in carbonates are not uniform; and the available database is insufficient.

To solve the problem it is necessary to use a comprehensive approach that includes the next procedures: (1) investigation of continuous sedimentary successions with carbonate beds and members; (2) age determination of carbonates based on reliable isotopic-geochronological data; (3) environmental analysis of studied deposits; (4) application of strict geochemical criteria to control the preservation state of isotopic systems in carbonates; and (5) selective leaching of noncogenetic carbonate phases from samples used in analysis. A fragment of seawater $^{87}Sr/^{86}Sr$ curve that characterizes time of the global Grenville orogeny and Rodinia emergence has been depicted, when this approach was applied to investigation of Middle-Upper Riphean boundary deposits in Siberia (Gorokhov *et al.*, 1995; Semikhatov *et al.*, 1998, 2002). This time span of geological record appears to be well understood now, whereas subsequent evolution of Sr isotope composition in the Late Riphean ocean is known to a first approximation (Kuznetsov *et al.*, 1997; Jacobsen and Kaufman, 1999; Walter *et al.*, 2000) and has been variably interpreted, because it is depicted mostly based on data for discrete intervals of sections in Arctic Canada and Spitsbergen, chronometry of which needs in verification.

Presenting in this work the new data on $^{87}Sr/^{86}Sr$ ratios in Upper Riphean carbonates from southern Urals, we also summarized and critically revised data of Sr isotope geochemistry, which characterize the key sections in Siberia, southern Urals, Australia, Arctic Canada, and Spitsbergen. We used the results to plot a consistent seawater $^{87}Sr/^{86}Sr$ curve and to discuss the geodynamic background of Sr isotope evolution in the Late Riphean ocean. New data are obtained for rocks of the Upper Riphean stratotype, the Karatau Group of southern Urals. A considerable stratigraphic range, abundance of slightly altered carbonate rocks, and representative biostratigraphic and isotopic-geochronological characterization are the obvious advantages of the studied sedimentary succession, which distinguish it from the others mentioned above.

STRATIGRAPHY, AGE LIMITS, AND STRUCTURE OF THE KARATAU GROUP

Thick (3.5 to 5.5 km) terrigenous-carbonate deposits of the Karatau Group are widespread in the western limb of the Bashkirian meganticlinorium, a large thrust-and-fold structure of the southwestern Urals. Deposits of the group discordantly overlie the Yurmatin Group (Middle Riphean stratotype) and have unconformable boundary with overlying terrigenous sediments of the Vendian Asha Group. Based on single isotopic dates and presence of tilloids, the lower part of the latter is attributed to the Lower Vendian (*Stratotip rifeya...*, 1983; Bekker, 1988; Maslov *et al.*, 2001, and references therein).

In the regional stratigraphic scheme that is adopted at present (*Stratotip rifeya...*, 1983; Kozlov *et al.*, 1991; *Unifitsirovannye...*, 1993; Maslov *et al.*, 2001), the Karatau Group is divided into six units (from the base upward): the siliciclastic Zilmerdak (1200 to 3300 m thick), variegated Katav (limestone beds 150–350 m thick), compositionally variable siliciclastic carbonate Inzer (300–800 m thick), limestone-dolomitic Min'yar (350–650 m), siliciclastic-carbonate Uk (160–350 m thick), and terrigenous Krivaya Luka (up to 560 m thick) formations. The Uk Formation crowns the group section in the western limb of Bashkirian meganticlinorium, and the last one is preserved after the pre-Vendian erosion in the east of the structure only. Three lower formations are interrelated, having gradual in-between transitions, whereas stratigraphic unconformities are recorded at the Uk Formation base and locally underneath the Min'yar Formation (Bekker, 1961, 1988; Raaben, 1975; Kozlov *et al.*, 1991; Maslov *et al.*, 2001; Maslov, 2002a). First results of C-isotope chemostratigraphy showed that the studied sections of the Min'yar Formation are lacking horizons with high positive $\delta^{13}C$ values, which are characteristic of the upper Upper Riphean in various continents (Kaufman and Knoll, 1995; Kaufman *et al.*, 1997; Kah *et al.*, 1999; Walter *et al.*, 2000). Seeking for explanation, Podkovyrov with colleagues (1998) and Raaben (2001) argued that some intervals of the section could be eroded in the pre-Uk time, although their assumption seemed doubtful, because upper Min'yar horizons are insufficiently exposed in the study region (Raaben, 2001). However, the Min'yar Formation is lacking the mentioned intervals not only in two sections described by Podkovyrov *et al.* (1998), but also in the section exposed along the Kuzhai Creek upstream of the Bakeevo Village, where it was sampled in detail and then studied in terms of C-isotope chemostratigraphy. In addition, the regional character of the pre-Uk hiatus is evident from certain

geological data: the contact between the Min'yar and Uk formations is locally marked by brecciated Fe–Mn accumulations, and basal sandstone beds of the Uk Formation, which enclose gravel and rare conglomerate interlayers, rest on different upper horizons (Shubino Member of limestones) of the Min'yar Formation and locally overlap the pre-Shubino dolostones (Bekker, 1961, 1988; *Stratotip rifeya...*, 1983; Maslov *et al.*, 2001; Raaben, 2000). Thus, we may regard as established a considerable scale of the pre-Uk unconformity. As for the hiatuses traceable at the base and in the middle of the Min'yar Formation, they have been observed in places thus signifying short-term events of subaerial exposition of local sedimentation settings.

Age limits of the Karatau Group are established based on some paleontological data and on isotopic dates of unequal validity. The isochron Pb–Pb date of 836 ± 25 Ma that has been obtained for limestones of the lower Inzer Subformation (Ovchinnikova *et al.*, 1998) is the most reliable one. It defines the time of early diagenesis in the lower Inzer limestones being consistent with the Rb–Sr dates ranging from 835 to 805 Ma, which characterize the authigenic 1M illites from the same subformation (Gorokhov *et al.*, 1995) and correspond to the age of burial diagenesis of sandy-clayey sediments underlying the lower Inzer limestones. The Pb–Pb age of dolostones from the middle part of the Min'yar Formation is 780 ± 85 Ma (Ovchinnikova *et al.*, 2000). A rather high uncertainty of this date is connected with an irregular style of epigenetic alterations in the rocks affected by sulfide mineralization. A reliable dating method has been applied as well to glauconites from the lower Uk Subformation. The Mössbauer spectroscopy showed that epigenetic alterations did not affect the structure of the glauconites, and consequently their isotopic-geochronological systems remain intact (Zaitseva *et al.*, 2000). The isochron Rb–Sr and K–Ar ages of these glauconites are 664 ± 11 and 669 ± 16 Ma, respectively. A close age value of 688 ± 10 Ma has been earlier obtained by the isochron Rb–Sr method for Al-glauconites from the lower Uk strata (Gorozhanin and Kutyavin, 1986). All these dates define the time of early diagenesis in the lower Uk sediments. The K–Ar dates obtained in the 1960s and 1070s for mineralogically unstudied glauconites from different horizons of the Karatau Group are less reliable. They represent a series of age values decreasing upward in the stratigraphic succession from 938 Ma in the upper Katav and lower Inzer strata to 896–853 in the middle and upper parts of the Inzer Formation, and then to 791–740 Ma in the upper member of the latter. The next range of 713–680 Ma characterizes the lower Min'yar Subformation, and final values of 658 to 630 Ma have been measured in the upper Uk Subformation (*Stratotip rifeya...*, 1983).

The upper age limit of the Karatau Group is defined by the isochron Rb–Sr date of 618 ± 13 Ma obtained for glauconite from the lower part of the Asha Group (Kozlov and Gorozhanin, 1993) and by the range of K–Ar dates (600 to 557 Ma) characterizing the mineralogically unstudied glauconites from overlying strata of the group (*Stratotip rifeya...*, 1983). There were also attempts to estimate age limits of the Karatau Group based on K–Ar dates for whole-rock samples of gabbro-diabases regarded as the pre- and post-Karatavian intrusions (*Stratotip rifeya...*, 1983, and other works). These attempts are not convincing, however: first, the K–Ar isotopic systems of magmatic whole-rocks samples are labile; and second, geological relations of the dated gabbro-diabase bodies with basal and terminal horizons of the group are inadequately known.

Paleontological constraints, first of all the data on succession of organic-walled microfossil assemblages in the Zilmerdak Formation (Veis *et al.*, 2000, 2003), which elucidate chronometric limits of deposits under consideration, are also important. In the lower part of the formation divided into the Biryan, Nugush, and Lemeza subformations, only the middle subdivision is bearing representative microbiota because of the facies control. In addition to small transit forms of simple morphology, the Nugush Subformation yields large and/or relatively complex taxa, such as *Chuaria, Navifusa, Simia nerujenica, Ostiana, Polyspheroides contexus, Asperatofilum, Taenitrichoides, Plicatidium, Rectia, Botuobia, Majaphyton, Lakhandinia, Caudiosphaera, Pseudodendron*, and some other morphotypes. Nearly a half of them is known in the Urals from the Kuzha and/or Tyulma sediments occurring below the Karatau Group (Veis *et al.*, 2000), and almost all the taxa are also characteristic of the overlying Bederysh Subformation that crowns section of the Zilmerdak Formation. Nevertheless, the Bederysh microbiota fundamentally differs from that of the Nugush Subformation, because it includes acanthomorphic acritarchs *Trachyhystrichosphaera aimica, T. stricta, T. parva, T. truncata* and associated *Tortunema, Glomovertella*, and *Arctacellularia*, all appearing for the first time at this level in the southern Urals (Veis *et al.*, 2003). In Siberia, acritarchs of the genus *Trachyhystrichosphaera* define the compositional aspect of the well-known Lakhanda microbiota from the Uchur-Maya region and of concurrent Derevnya microbiota from the Turukhansk Uplift (German, 1990; Veis *et al.*, 1998, 2000). According to available dates obtained for carbonate rocks by the isochron Pb–Pb method, the maximum age limit of these biotas is 1030 Ma (Semikhatov *et al.*, 2000, 2003; Ovchinnikova *et al.*, 2001). The assemblage of organic-walled microfossils from underlying horizons of Siberian sections mentioned above, i.e., from the Kerpyl Group of the Uchur-Maya region and its analogues in the Turukhansk Uplift, which accumulated 1200–1030 Ma ago (Bartley *et al.*, 2001; Semikhatov *et al.*, 2002), is taxonomically different. It includes a set of taxa characteristic of the Nugush Subformation of the Zilmerdak Formation and of the Kuzha and Tyulma sediments in the southern Urals (Veis *et al.*, 1998, 2000, 2003). Consequently, the pre-Bederysh level in the Uralian succession of organic-

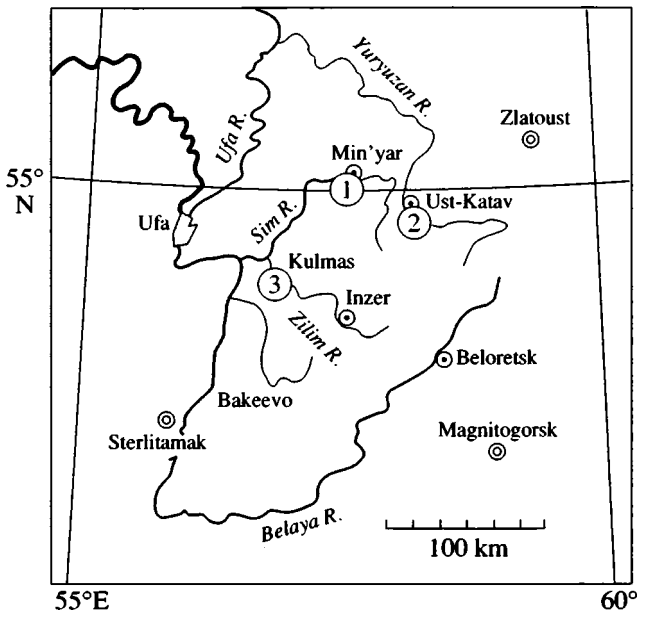

Fig. 1. Localities of the Min'yar (1), Shubino (2), and Kulmas (3) sections of the Karatau Group.

walled microbiotas corresponds to the pre-Lakhanda level in Siberia, and the Bederysh Subformation of the Zilmerdak Formation is not older than 1030 Ma. On the other hand, silicified microfossils occurring in the Karatau Group above the latter formation are represented by transit taxa (Sergeev, 1992), whereas a specific assemblage of organic-walled forms is recorded here in the lower part of the Inzer Formation only (Veis *et al.*, 2000, 2003). This one is similar to the assemblage from middle horizon of the Akademikerbreen Group of Spitsbergen (Butterfield *et al.*, 1994), but their similarity is unimportant for chronometric calibration of the Karatau Group, because the former group has not been dated by means of isotopic methods.

It is important for the current consideration that *Melanocerillium* remains are reported to occur in the middle of the Min'yar Formation (Maslov *et al.*, 1994), though they have not been illustrated and described to our regret. Representatives of this genus attributed to testate amoebae, are known from a series of Upper Riphean sections in North America, West Europe and South Asia, and they appear in the geological record just before 800 Ma (Porter and Knoll, 2000). Thus, the verification of *Melanocerillium* occurrence in the Min'yar Formation should support evaluation of Pb–Pb age of this formation (780 ± 85 Ma).

If the linear dependence between thickness and accumulation time of Karatavian deposits is admissible, then, using the presented radiometric dates and some results of chemostratigraphy considered below, we can estimate the time spans for subdivisions of the group. The estimated ranges are 890–850 Ma for the Katav Formation, 850–810 Ma for the Inzer Formation (850 to 830 and 830 to 810 Ma for its lower and upper subformations, respectively), 810–770 Ma for the Min'yar Formation, and 690–640 Ma for the Uk Formation divided into the lower (690–660 Ma) and upper (660–640 Ma) subformations. These ranges are consistent with data on stromatolite succession in the Karatau Group (Semikhatov and Raaben, 1994, 1996) and do not contradict the sedimentation rate assessments for Upper Riphean isofacies in the Urals (15 to 20 m/m.y., Maslov *et al.*, 2002b) and Siberia (15 to 30 m/m.y., Semikhatov *et al.*, 2000, 2002; Ovchinnikova *et al.*, 2001). The considerable difference between isotopic dates obtained for the Min'yar and Uk formations confirms a long pre-Uk hiatus.

In the western limb of the Bashkirian meganticlinorium, Karatavian deposits of inner and outer facies zones are separated by the Zilmerdak thrust and associated faults (*Stratotip rifeya...*, 1983). The zones differ from each other mostly in abundance of terrigenous sediments above the Zilmerdak Formation: up to 50% in the inner (eastern) zone and only 10 to 15% in the outer (western) zone. The degree of epigenetic alterations in sediments decreases in the same direction. We studied three sections of the outer zone characterizing predominantly the carbonate part of the Karatau Group. One is exposed near the town of Min'yar, second one along the Yuryuzan River in the Ust-Katav area upstream of the Shubino Village, and the third section corresponds to a series of road cuts along the Ufa-Beloretsk highway, which are situated near the Kulmas Settlement (Fig. 1). Further, they are termed as the Min'yar, Shubino, and Kulmas sections. In the first one, we studied sediments of the Katav, Inzer, and Min'yar (lower part) formations. Exposed in the second section are rocks of the Min'yar (upper part) and Uk formations, whereas the road cuts exhibit fragments of the Katav, Inzer, and upper Uk successions. Sediments of the formations are described below, and stratigraphic positions of studied samples are shown in Fig. 2.

The Katav Formation conformably resting on siliciclastic Zilmerdak deposits is divided in two subformations. In the Min'yar section, the lower subformation (120 m) is composed of cherry-red flaggy clayey limestones (10 to 24% of siliciclastic components) with rare marl and calcareous shale interlayers, the latter concentrated near the base. Gently inclined cross lamination characteristic of lower strata grades upward into fine horizontal stratification. The upper subformation (60 m) is represented by alternating pale-yellow to pinkish and greenish gray flaggy clayey limestones (10 to 15% of siliciclastic components), characteristic of which is banding and microbial lamination in the upper part. Sediments enclose rare flakestone lentils and slightly dolomitized (Mg/Ca = 0.054) in some members. Micritic matrix of limestones encloses microscopic lenses of microsparite and less abundant segregations (of mm-size scale) and veinlets of fine-grained dolosparite. Cathodoluminescence of micrite and microsparite is weak to moderate, variegated from yel-

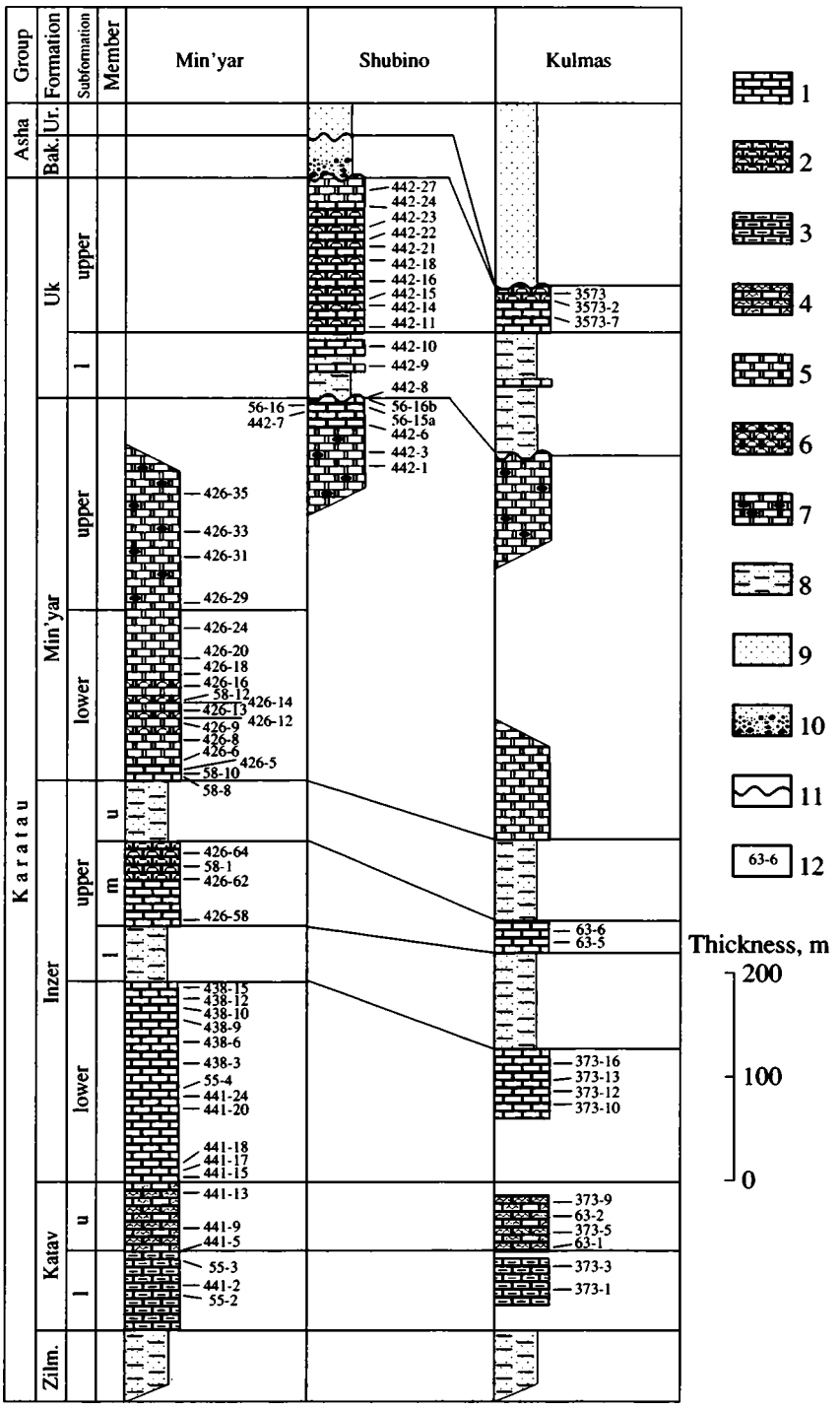

Fig. 2. Lithostratigraphy of the Karatau Group, the Min'yar, Shubino, and Kulmas sections: (1) limestone; (2) stromatolitic limestone; (3) clayey limestone; (4) microbial striate limestone; (5) dolostone; (6) stromatolitic dolostone; (7) dolostone with cherty concretions; (8) sandy-clayey deposits; (9) sandstone; (10) sandstone with gravelstone interlayers; (11) stratigraphic unconformity; (12) sample nos. and sampling levels. Abbreviations: (Bak.) Bakeevo; (Zilm.) Zilmerdak; (Ur.) Uryuk; (l) lower; (m) middle; (u) upper.

low to pinkish; it is of the same type though brighter in dolosparite segregations. Fragments of Katav sediments exposed in the Kulmas section do not differ in primary lithology, but the rocks are more intensely fractured here and impregnated by calcite veinlets of late generation.

The Inzer Formation is also divided in two subformation, the lower of which was known for a while under the name "Podinzer Beds." In the Min'yar section, the lower subformation (ca. 190 m) is represented by dark gray to black limestone weakly dolomitized (Mg up to 1.1%, Mg/Ca = 0.030) in the middle 35-m-

thick member. Packages of stromatolitic limestones and flakestone lentils are recognizable in the upper part. In the lower 25 to 30-m-thick interval, recrystallization of rock is somewhat stronger than in the Katav Formation. Vaguely laminated micrite with thin microsparite lenses and patchy dolomitization responsible for development of rhombohedral dolomite crystals are established in this interval. However, the main part of the lower Inzer limestones is insignificantly recrystallized, being composed of alternating fine micrite and microsparite laminae, among which there are distinguishable only microscopic lenses of microcrystalline secondary calcite. Microsparite laminae (1 to 3 mm thick) retain the relic clastic structure. Content of siliciclastic fraction is commonly less than 1%, and rare interlayers of clayey limestone (up to 9.6% of siliciclastic components) resembling the Katav Formation rocks are established near the unit base only (Table 1). In distinction from the Min'yar section, rocks in the middle interval of lower Inzer deposits that is fragmentary exposed in the Kulmas section have microcrystalline texture and reveal a system of fine (0.1 to 1.0 mm) fractures filled with late crystalline calcite and presence of microstylolite seams. Limestone sample 373-16 collected near the apparent top of the Kulmas section is of a mosaic texture, with dispersed inclusions of Fe-hydroxides.

In the Min'yar section, two siliciclastic (gray feldspar-quartz-glauconite sandstones, siltstones and shales) members (70 and 50 m) of the upper Inzer Subformation are separated from each other by the middle carbonate member (80 to 120 m). Rocks of the latter are massive microphytolitic, fine-laminated micritic, stromatolitic, and clastic limestones; two latter varieties appear closer to the subformation top. Fine siliciclastic admixture in the rocks (usually less than 3%) may be concentrated sometimes to a greater extent in some basal horizons. In the Kulmas section, where the succession and lithology of members is unchanged, the total thickness of subformation is decreased (180 m) owing to the carbonate member reduction (Fig. 2). A weak cathodoluminescence is characteristic of limestones in the lower 25 to 30-m-thick part of the Inzer Formation. In other rocks, the bright pinkish to yellow or moderate yellow luminescence is characteristic of the late calcite and dolomite, respectively.

The Min'yar Formation is the thickest (360 to 400 m in the studied sections) and entirely carbonate subdivision of the Karatau Group. In the synonymous section, it is predominantly composed of light to pinkish gray clastic fine-laminated and massive stromatolitic dolostones, which are sandwiched between the basal (12 to 14 m) and terminal Shubino (15 to 20 m) members. In the upper formation interval, the appearance level of dark gray to black lenses and interlayers of early diagenetic cherts is adopted for the boundary between the lower Minka (170–180 m) and upper Byanka (160–180 m) subformations. Near the Minka Subformation base, there is a marker dolostone bed with stromatolites *Conophyton* that has been deposited below the storm wave base. In the middle of that subformation, there are some bedding planes marked by desiccation cracks.

Characteristic of the Min'yar dolostones are the narrow range of Mg/Ca ratios (from 0.552 to 0.636) and low content of siliciclastic components (3% in average). Only within the interval from 55 to 90 m above the formation base, percentage of siliciclastic fraction grows up to 17–40%. Stromatolitic varieties of rocks correspond to dolomicrites and dolosparite, and the clastic ones are of the dolosparite type. The rocks retain their primary structures, but they may contain the late crystalline dolomite (5%) filling in the small caverns or thin (0.5 to 2 mm) lenses and veinlets. Dolomite of this generation is mostly localized in the lower 70 to 80-m-thick interval of the formation section.

Cherty segregations in the Min'yar dolostones yield silicified microfossils, some of which are in the extraordinary preservation state (Sergeev, 1992). Consequently, the segregations were formed at the earliest stage of diagenesis, immediately after posthumous burial of microorganisms in bottom-surface deposits near the water-sediment interface (Maliva *et al.*, 1989; Sergeev, 1992). Dolomicrite relicts present in some cherty concretions suggest that the early dolomitization of the Min'yar carbonates also begun immediately after precipitation of sedimentary material that likely corresponded in composition to high-Mg calcite. Limestones from the formation lower member are slightly dolomitized (Mg/Ca up to 0.035), composed of micrite and microsparite with microscopic accumulations (0.1 to 0.3 mm across) of rounded micritic clasts. They bear 5 to 10% of siliciclastic admixture.

In the Min'yar section, we sampled the Minka to lower Byanka deposits that are 350 m thick and well exposed. Upper horizons of the formation, i.e., the upper Byanka dolostones and Shubino limestones, have been studied in the Shubino section. The Byanka dolostones exposed here are identical to those of the Min'yar section. Characteristic of them is moderate red, sometimes patchy cathodoluminescence. Limestones of the Shubino Member usually consist of pure calcite, and a weak enrichment in Mg (Mg/Ca up to 0.16) is established for them only near the contact with underlying dolostones. These limestones with sporadic desiccation cracks bear an insignificant admixture of clastic quartz (0.5 to 3%), and their micrite to microsparite matrix encloses small accumulations of subangular micritic clasts. Being replaced to some extent by the mosaic sparite or coarse calcite crystals, the latter are surrounded by thin rims with radiolitic texture and/or by Fe-hydroxides. There are also small veins and nests of coarse calcite crystals. In contrast to non-luminescent micrite and microsparite, luminescence of this calcite is bright.

The Uk Formation consists of two subformations. The lower subformation that is 70 m thick in the Shu-

Table 1. Trace element concentrations and characteristics of siliciclastic admixture in carbonate rocks of the Karatau Group, southern Urals

Sample no.[1]	Section[2]	Sampling level[3], m	Rock type[4]	Siliciclastic admixture (SA),[5] %	Mg, %	Mn, ppm	Fe, ppm	Sr, ppm	Mg/Ca	Mn/Sr	Fe/Sr	Minerals of SA[6]
Uk Formation												
442-27	SH	195	D	1.8	9.5	110	1120	12	0.350	9.33	93.3	–
442-24	SH	180	L	0.9	1.5	120	220	53	0.040	2.28	4.2	–
442-23	*SH*	*160*	*L*	*0.6*	*0.5*	*22*	*130*	*313*	*0.014*	*0.07*	*0.4*	–
442-22	SH	150	L	2.9	0.5	31	280	148	0.014	0.21	1.9	–
442-21	SH	140	L	3.5	0.3	33	340	109	0.008	0.30	3.1	–
442-18	SH	125	L	3.8	0.3	34	250	106	0.008	0.32	2.4	–
442-16	SH	105	L	6.0	1.1	150	810	117	0.028	1.28	6.9	–
442-15	SH	90	L	11.4	0.4	80	770	140	0.010	0.56	5.5	Q, Fsp, Mc, Chl
442-14	SH	85	L	13.6	0.3	82	1010	110	0.007	0.75	9.2	Q, Fsp, Mc, Chl
3573	K	84	L	0.8	0.2	66	410	90	0.006	0.73	4.6	–
3573-2	K	78	L	1.1	0.1	64	260	146	0.002	0.44	1.8	–
3573-7	*K*	*62*	*L*	*3.1*	*0.3*	*27*	*470*	*437*	*0.008*	*0.06*	*1.1*	–
442-11	*SH*	*60*	*L*	*1.0*	*0.1*	*100*	*1120*	*715*	*0.002*	*0.14*	*1.6*	–
442-10	SH	45	L	8.1	0.4	603	3450	289	0.010	2.09	11.9	Q, Fsp
442-9	SH	25	L	5.0	0.4	220	1700	281	0.011	0.78	6.1	
Min'yar Formation												
442-8	SH	362	L	1.6	0.8	385	1420	340	0.019	1.13	4.2	–
56-16b	SH	361	L	0.3	0.3	130	150	477	0.008	0.27	0.3	Q
56-16	SH	360	L	2.6	0.1	535	380	346	0.002	1.55	1.1	–
56-15a	SH	359	L	3.2	0.1	445	670	375	0.002	1.18	1.8	Q
442-7	SH	358	L	3.4	0.3	1460	660	404	0.007	3.61	1.6	Q, Fsp
442-6	SH	343	L	1.6	5.1	2960	1420	230	0.156	12.9	6.2	–
442-3	SH	315	D	3.2	13.4	340	1340	20	0.604	17.1	67.1	–
442-1	*SH*	*300*	*D*	*2.1*	*13.4*	*102*	*440*	*113*	*0.608*	*0.90*	*3.9*	–
426-35	M	289	D	2.2	13.4	96	240	68	0.609	1.41	3.5	–
426-33	*M*	*238*	*D*	*0.7*	*13.3*	*63*	*150*	*103*	*0.636*	*0.61*	*1.5*	–
426-31	M	213	D	1.6	13.2	68	90	44	0.612	1.55	2.1	–
426-29	M	170	D	8.9	13.2	230	320	28	0.608	8.21	11.4	Q, Mc, Chl, FO
426-24	*M*	*144*	*D*	*0.9*	*13.4*	*51*	*102*	*70*	*0.612*	*0.73*	*1.4*	–
426-20	M	116	D	2.6	13.4	101	150	75	0.606	1.35	2.0	–
426-18	*M*	*102*	*D*	*1.8*	*13.2*	*44*	*140*	*55*	*0.608*	*0.80*	*2.6*	–
426-16	M	87	D	17.0	12.2	99	500	57	0.552	1.74	8.7	Q, Fsp, FO
58-12[7]	M	75	D	0.1	12.7	102	180	44	0.557	2.31	4.1	–
426-14	M	72	D	4.1	13.2	115	220	54	0.601	2.13	4.1	–
426-13	M	65	D	6.6	12.7	98	360	13	0.593	7.54	27.7	Q, Fsp, FO
426-12	M	58	D	40.1	12.7	91	410	65	0.575	1.40	6.3	Q, Fsp
426-9	M	54	D	0.7	13.2	140	60	14	0.594	10.1	4.1	–
426-8	M	36	D	0.5	13.3	96	620	28	0.604	3.43	22.1	–
426-6	M	18	D	4.8	13.3	184	450	10	0.607	18.4	45.0	–
426-5	M	6	L	4.8	1.4	138	1370	195	0.035	0.70	7.1	Q, Fsp, Chl
58-10	M	4	L	4.7	0.4	125	1470	265	0.010	0.48	5.5	–
58-8	M	3	L	10.0	0.3	130	1670	183	0.009	0.73	9.1	Q, Fsp, Mc, Chl

Table 1. (Contd.)

Sample no.[1]	Section[2]	Sampling level[3], m	Rock type[4]	Siliciclastic admixture (SA)[5], %	Mg, %	Mn, ppm	Fe, ppm	Sr, ppm	Mg/Ca	Mn/Sr	Fe/Sr	Minerals of SA[6]
Inzer Formation												
426-64	*M*	*308*	*L*	*1.7*	*0.5*	*32*	*390*	*175*	*0.013*	*0.18*	*2.2*	Q
58-1	M	295	L	1.4	1.4	230	1820	118	0.038	1.97	15.4	Q
426-62	M	285	L	1.3	0.6	45	900	169	0.016	0.27	5.3	–
63-6	*K*	*270*	*L*	*1.8*	*0.2*	*43*	*495*	*243*	*0.006*	*0.17*	*2.0*	–
63-5	*K*	*260*	*L*	*2.4*	*0.8*	*38*	*570*	*213*	*0.020*	*0.18*	*2.7*	–
426-58	M	245	L	7.0	0.1	51	1510	298	0.002	0.17	5.1	Chl, Mc, Q, Fsp
438-15	*M*	*188*	*L*	*0.3*	*0.2*	*22*	*315*	*545*	*0.004*	*0.04*	*0.6*	–
438-12	*M*	*172*	*L*	*0.3*	*0.2*	*21*	*310*	*605*	*0.005*	*0.03*	*0.5*	–
438-10	*M*	*158*	*L*	*0.4*	*0.1*	*15*	*370*	*490*	*0.002*	*0.03*	*0.8*	–
438-9	*M*	*151*	*L*	*0.7*	*0.2*	*21*	*325*	*521*	*0.006*	*0.04*	*0.6*	–
438-6	*M*	*130*	*L*	*0.2*	*0.3*	*19*	*220*	*512*	*0.006*	*0.04*	*0.4*	–
438-3	*M*	*110*	*L*	*0.8*	*0.4*	*23*	*420*	*495*	*0.009*	*0.05*	*0.8*	–
373-16	K	109	L	3.5	0.4	94	810	151	0.010	0.62	5.4	Q
373-13	*K*	*90*	*L*	*4.4*	*0.4*	*11*	*160*	*590*	*0.010*	*0.02*	*0.3*	Q
373-12	*K*	*85*	*L*	*0.6*	*0.5*	*14*	*400*	*417*	*0.013*	*0.03*	*0.9*	–
55-4[7]	M	82	L	0.1	0.2	86	646	77	0.004	1.12	8.4	–
441-24	M	80	L	1.7	1.1	37	870	254	0.030	0.15	3.4	Q
373-10	*K*	*70*	*L*	*0.6*	*0.1*	*18*	*460*	*439*	*0.003*	*0.04*	*1.1*	–
441-20	*M*	*67*	*L*	*0.5*	*0.1*	*15*	*330*	*423*	*0.002*	*0.04*	*0.8*	–
441-18	*M*	*15*	*L*	*0.8*	*0.1*	*44*	*420*	*411*	*0.003*	*0.11*	*1.1*	–
441-17	M	9	L	4.6	0.1	96	2630	338	0.003	0.28	7.8	Q, Fsp
441-15	M	1	L	9.6	0.2	102	2640	508	0.005	0.20	5.2	Q, Fsp, Mc
Katav Formation												
441-13	M	130	L	10.2	0.4	270	3470	152	0.010	1.78	22.8	Q, Fsp, Mc, Chl
373-9	K	120	L	13.2	0.2	180	3870	240	0.006	0.75	16.1	Q, Mc, Chl, Fsp
63-2[7]	K	105	L	0.4	0.1	225	1210	345	0.002	0.65	3.5	–
441-9	M	95	L	13.5	0.1	220	4360	190	0.002	1.16	22.9	Q, Fsp, Mc, Chl
373-5	K	90	L	13.0	0.6	170	3630	215	0.015	0.79	16.8	Q, Fsp, Mc, Chl
63-1[7]	K	80	L	0.2	0.1	320	2970	255	0.002	1.25	11.6	–
441-5	M	70	L	15.1	0.1	240	3970	126	0.002	1.90	31.5	Q, Mc, Chl, Fsp
55-3[7]	M	65	L	0.5	0.1	324	3070	155	0.002	2.09	19.8	–
373-3	K	60	L	18.6	0.3	175	2480	160	0.007	1.09	15.5	Mc, Chl, Q, Fsp
441-2	M	40	L	10.5	0.1	234	1560	154	0.002	1.52	10.1	Q, Mc, Chl, Fsp
55-2[7]	M	35	L	0.2	0.1	420	2025	67	0.002	6.30	30.2	–
373-1	K	30	L	23.8	0.1	160	2220	120	0.004	1.33	18.5	Q, Fsp, Mc, Chl

Note: (1) Italicized are data for samples satisfying the accepted geochemical criteria (Mg/Ca ≤ 0.024, Fe/Sr ≤ 5.0, Mn/Sr ≤ 0.2 for limestones and Mg/Ca ≥ 0.608, Mn/Sr ≥ 1.2, Fe/Sr ≥ 3.0 for dolostones). (2) Sections by names: (M) Min'yar, (K) Kulmas, (SH) Shubino. (3) Sampling levels are denoted in meters above the formation base in the generalized succession. (4) Rocks: (L) limestone; (D) dolomite; (5) siliciclastic admixture; (6) Minerals: (Q) quartz, (Mc) mica, (Fsp) Feldspar, (Chl) chlorite, (FO) Fe-oxides, all listed in order of decreasing abundance. (7) Samples of veined calcite or dolomite.

bino section is composed of greenish gray glauconite-quartz sandstone, siltstone, and rare shale beds intercalated with separate interlayers (0.2 to 0.4 m) and two members (4–6 and 8–9 m) of micritic limestones, the latter confined to the upper part. Siliciclastic components of limestones (5 to 8%) are represented by quartz and clay minerals. The extent of dolomitization in limestones (Mg/Ca up to 0.04) is variable. Gravelstone at the base of subformation contains clasts of underlying carbonate rocks. The lower 120-m-thick interval of the upper Uk Subformation (140 m) is composed of light gray stromatolitic and rare clastic limestones, which are weakly dolomitized in contrast to dolostones and dolomitic limestones characteristic of the upper 20 to 25-m-thick interval. Micrite and microsparite with small sparite segregations are constituents of stromatolitic rock varieties, whereas clastic limestones are predominantly composed of micrite. In some rare rocks, micrite and microsparite are partially replaced by sparite and by medium-grained poikilitic calcite, and dispersed Fe-oxides and hydroxides appear in their matrix. Small rhombohedral dolomite crystals irregularly distributed in the matrix are surrounded by ferruginate rims. The Mg concentration in limestones from the lower 120-m-thick interval of the upper Uk Subformation ranges from 0.2 to 1.5%, and percentage of siliciclastic admixture is usually not greater than 0.1–6%, growing up to 11–14% only near the subformation base. In dolostones and dolomitic limestones crowning the subformation section and containing only 1–2% of siliciclastic fraction, the Mg content ranges from 1.5 to 9.5%. These rocks are composed of dolosparite matrix with irregular segregations and veinlets of coarse-crystalline dolomite, in which there are microscopic micrite relics (0.1 to 0.3 mm) and corroded crystals of poikilitic calcite. The rhombohedral dolomite crystals have regeneration rims and are zonal sometimes. In the Kulmas section, where the Uk Formation is of the same structure in general, its lower subformation increases in thickness up to 150 m and encloses only one limestone package 5 m thick (Fig. 2). In contrast, the upper subformation is thinner (40 m) than in the Shubino section and enriched in microphytolitic limestones. A considerable portion of the upper subformation is likely eliminated by the pre-Vendian erosion. According to cathodoluminescence analysis, the upper Uk carbonates are inhomogeneous. In the non-luminescent micrite to microsparite matrix of limestones, there are seen luminescence of variable intensity that is characteristic of sparite segregations and veins. The patchy luminescence of dolostones indicates that they are composed of several carbonates' generations.

The predominantly carbonate units of the Karatau Group, which are described above, accumulated in a vast sea basin, where, according to conclusion of Maslov *et al.*, (2001), relatively deep environments located below storm wave base were subordinate to dominant subtidal settings. The analyzed structural peculiarities of rocks, widespread among which are flakestones, and the data on stromatolite morphotypes suggest that the Katav, lower Inzer, and lower Minka carbonates mostly represent sediments of a lower subtidal zone situated close to the storm wave base. Stromatolitic horizons present in these units and in the Uk Formation correspond to the open-shelf deposits. The shallowest origin of carbonates, as compared to others of the Karatau Group, is inferable for their microphytolitic varieties occurring in upper horizons of the Inzer and Uk formations and for those packages of the upper Minka, Byanka and Uk successions, which reveal marks of subaerial exposition.

INVESTIGATION PROCEDURE

Among 250 samples of carbonate rocks collected from the Karatau Group, we selected for investigation of Rb–Sr systematics only those 70 samples, which were macroscopically homogeneous, lacking obvious signs of secondary recrystallization (53 limestone and 17 dolostone samples; Table 1, Fig. 2). In addition, we analyzed samples of veined calcite from various horizons of the Katav Formation and lower Inzer Subformation and of vein dolomite from the Min'yar Formation. Selected samples were sawed in two parts, one of which was examined by petrographic and cathodoluminescence methods, and the other one was used for the X-ray, chemical, and isotopic analyses. Ca and Mg contents in carbonate fractions were measured by wet chemical method, and the atomic absorption spectroscopy was applied to determine the Mn and Fe concentrations. Mineral composition of siliciclastic admixture is defined by the X-ray diffraction method.

In order to prepare samples for the Rb–Sr isotopic analysis, we removed their epigenetic carbonate phases applying the procedure of selective dissolution that included the preliminary treatment of powdered material in solution of ammonium acetate (NH_4OAc) and subsequent dissolution of residue in acetic acid as it is described in work by Gorokhov *et al.* (1995). After addition of 3 ml of 1N NH_4OAc, we subjected each sample weighing to ultrasonic machining in UZDN-2T device during 10 minutes and repeated the operation for the residue after centrifuging. After second centrifuging, the precipitate was treated in 10% acetic acid. The carbonate components of limestones dissolved completely after 72 hours, and those of dolostones entered the solution after 72 to 120 hours. Next, the insoluble silicate fraction was separated by centrifuging and rinsed two times in the twice-distilled water. Centrifugates of two treatment stages are designated in this work as the AMA (ammonium acetate) and ACA (acetic acid) phases.

Effectiveness of preliminary sample treatment in ammonium acetate has been demonstrated in a series of works devoted to Rb–Sr systematics of Proterozoic and Phanerozoic carbonates (Gorokhov *et al.*, 1995; Kuznetsov, 1998; Bailey, 2000; Bartley *et al.*, 2001; Semikhatov *et al.*, 2002). This procedure facilitates a partial

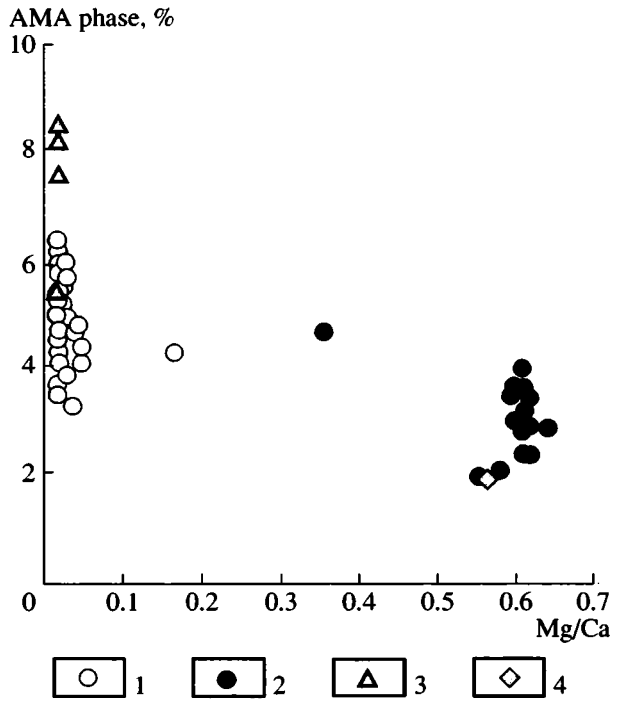

Fig. 3. Diagram showing percentage of AMA phase versus Mg/Ca ratios in the studied samples: (1) limestone; (2) dolostone; (3) veined calcite; (4) veined dolomite.

removal of secondary carbonate phases thus leading to enrichment of samples in primary material. The AMA phase leached from surface of mineral particles represents the latest epigenetic generation of carbonate minerals, whereas the ACA phase corresponds mostly to primary carbonate material with a considerably lesser admixture of epigenetic phases. The completeness of separation and a possibility-in-itself to separate two phases using the selective dissolution depend on degree of epigenetic alterations in primary carbonate sediment.

The Rb and Sr concentrations in AMA and ACA phases are measured using the isotopic dilution method and mixed ^{87}Rb–^{84}Sr spike. Sr isotope composition was determined on the multi-collector Finnigan MAT 261 mass spectrometer using a regime of simultaneous measurements of all isotope currents. The laboratory averages of 0.71025 ± 0.00001 ($2\sigma_{av.}$, n = 38) and 0.70917 ± 0.00002 ($2\sigma_{av.}$, n = 2) were determined for the NIST SRM 987 and USGS EN-1 standard carbonates, respectively, during the period of data gathering.

Rb–Sr SYSTEMATICS OF NON-COGENETIC CARBONATE PHASES

Percentages of AMA phases range from 3.2 to 6.4% (4.6% in average) in the Karatavian limestones and from 2.0 to 4.7% (3.1% in average) in dolostones. The highest percentage (7.6 to 8.6%) is characteristic of veined calcite, whereas the lowest values are determined in vein dolomite (2.0%) and in dolostones (2.0 to 2.1%) enriched in siliciclastic material. In general, the content of AMA phase decreases parallel to the Mg/Ca ratio growth (Fig. 3). The ACA and AMA phases differ from each other in terms of Rb–Sr characteristics (Table 2). Concentration of Rb in AMA phases of limestones and dolostones is 34 and 50 times greater, respectively, than in corresponding ACA phases. The Sr concentration is almost equal in both phases of limestones, whereas in dolostones it is two times greater in the AMA than in the ACA phase. This behavior of trace elements formerly established in Riphean carbonates of Siberia implies that the AMA phase is represented in both rock types by secondary calcite and/or by low-Mg dolomite (Gorokhov *et al.*, 1995; Semikhatov *et al.*, 2002). The Sr concentration in bulk carbonate material (Table 1) used to calculate Mn/Sr and Fe/Sr ratios has been assessed as the weighted average based on this element concentrations in AMA and ACA phases and on percentages of the latter in the rocks (Table 2).

The $^{87}Sr/^{86}Sr$ ratios measured in AMA phases of limestones and dolomites are usually higher than in ACA phases. Difference in this parameter between two phases under consideration is up to 0.0146 in dolostone and not greater than 0.0080 in limestones (Table 2, Fig. 4). The inversed relations are characteristic of veined calcite, and $^{87}Sr/^{86}Sr$ ratios in ACA phases of this mineral exceed those in AMA phases by values of 0.0001 to 0.0020 (Fig. 4a). All the mentioned distinctions evidence the non-cogenetic character of ACA and AMA phases. In addition, difference between $^{87}Sr/^{86}Sr$ ratios in these phases of limestones lacking signs of dolomitization (an exception is sample 442-7 taken in proximity to dolostones) is positively correlative with Mn/Sr (correlation coefficient r = 0.67) and Fe/Sr (r = 0.88) ratios. This peculiarity means that carbonate rocks became enriched in Mn, Fe, and radiogenic ^{87}Sr during their partial recrystallization at the late stages of lithogenesis. Even the late veined calcite was partially recrystallized under influence of epigenetic fluids, because its phases are non-cogenetic. However, none of the correlations mentioned above has been established for dolostones, which may indicate deeper recrystallization of these rocks.

Data considered above imply that Rb–Sr systems of ACA and AMA phases in limestones and dolostones have been formed at different stages of lithogenesis. Accordingly, it would be incorrect to calculate the $^{87}Sr/^{86}Sr$ initial ratios for AMA phases based on the sample age and Rb and Sr concentrations in the bulk carbonate material, because they may turn out to be significantly lower than ratios obtained for the paired ACA phases (Table 2, Fig. 4). Preliminary leaching of samples in ammonium acetate leads to a more confident assessment of $^{87}Sr/^{86}Sr$ ratios in a paleobasin, as they are estimated in this case based on analytical results obtained for materials enriched in primary carbonates. Thus, all the values of $^{87}Sr/^{86}Sr$ ratios presented below characterize the studied ACA phases only, if the other

Table 2. Rb–Sr isotopic data characterizing carbonate phases of limestones and dolostones of the Karatau Group

Sample no.[1]	Rock[2]	Carbonate phase[3]	Content of the phase (%)	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ measured	$^{87}Sr/^{86}Sr$ initial
Uk Formation								
442-27	D	AMA	4.7	3.38	96	–	–	–
		ACA	94.0	0.08	7.7	0.0313	0.71080	0.71050
442-24	L	AMA	4.4	1.00	67	0.0429	0.71239	0.71199
		ACA	94.3	0.02	52	0.0010	0.71188	0.71187
442-23	*L*	*AMA*	*4.8*	*1.73*	*343*	*0.0146*	*0.70579*	*0.70565*
		ACA	*94.4*	*0.03*	*311*	*0.0003*	*0.70545*	*0.70545*
442-22	L	AMA	4.5	3.85	190	0.0585	0.70756	0.70702
		ACA	92.4	0.19	148	0.0037	0.70598	0.70595
442-21	L	AMA	5.3	9.51	112	0.1273	0.70785	0.70667
		ACA	91.3	0.13	111	0.0034	0.70649	0.70646
442-18	L	AMA	4.4	4.77	121	0.1138	0.70933	0.70827
		ACA	92.6	0.10	105	0.0028	0.70650	0.70647
442-16	L	AMA	3.2	13.0	153	0.2460	0.71119	0.70891
		ACA	91.3	0.20	116	0.0050	0.70725	0.70720
442-15	L	AMA	3.7	9.75	213	0.1325	0.71107	0.70984
		ACA	85.3	0.18	138	0.0037	0.70745	0.70742
442-14	L	AMA	3.7	19.7	48	–	–	–
		ACA	82.3	0.28	118	0.0069	0.70806	0.70800
3573	L	AMA	5.4	5.44	116	0.1362	0.70741	0.70615
		ACA	93.9	0.17	88	0.0056	0.70596	0.70591
3573-2	L	AMA	5.1	4.03	176	0.0662	0.70696	0.70635
		ACA	94.0	0.05	144	0.0010	0.70599	0.70598
3573-7	*L*	*AMA*	*5.8*	*3.07*	*522*	*0.0170*	*0.70578*	*0.70562*
		ACA	*90.9*	*0.06*	*431*	*0.0004*	*0.70538*	*0.70538*
442-11	*L*	*AMA*	*4.6*	*2.19*	*771*	*0.0082*	*0.70612*	*0.70604*
		ACA	*94.2*	*0.06*	*712*	*0.0002*	*0.70581*	*0.70580*
442-10	L	AMA	5.3	9.29	177	0.1522	0.70807	0.70666
		ACA	86.6	0.15	296	0.0015	0.70596	0.70595
442-9	L	AMA	4.1	3.82	218	0.0506	0.70694	0.70647
		ACA	91.4	0.10	282	0.0010	0.70583	0.70582
Min'yar Formation								
442-8	L	AMA	5.7	1.59	410	0.0114	0.70700	0.70689
		ACA	92.7	0.21	335	0.0017	0.70660	0.70658
56-16b	L	AMA	4.3	0.53	465	0.0330	0.70629	0.70626
		ACA	95.4	0.05	478	0.0003	0.70621	0.70621
56-16	L	AMA	4.3	3.25	464	0.0203	0.70654	0.70634
		ACA	93.4	0.06	342	0.0005	0.70608	0.70607
56-15a	L	AMA	4.8	4.49	463	0.0280	0.70652	0.70624
		ACA	93.6	0.07	372	0.0005	0.70601	0.70600
442-7	L	AMA	4.4	4.52	447	0.0262	0.70653	0.70627
		ACA	92.2	0.08	404	0.0005	0.70608	0.70607
442-6	L	AMA	4.3	2.37	205	0.0308	0.70701	0.70671
		ACA	94.0	0.23	230	0.0030	0.70624	0.70621

Table 2. (Contd.)

Sample no.[1]	Rock[2]	Carbonate phase[3]	Content of the phase (%)	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ measured	$^{87}Sr/^{86}Sr$ initial
442-3	D	AMA	2.8	2.09	24	0.1720	0.70816	0.70643
		ACA	94.0	0.03	20	0.0033	0.70682	0.70678
442-1	*D*	*AMA*	*2.4*	*6.60*	*715*	*0.0270*	*0.70784*	*0.70754*
		ACA	*90.9*	*0.06*	*101*	*0.0017*	*0.70613*	*0.70611*
		AMA[5]	*3.3*	*4.10*	*209*	*0.0575*	*0.70766*	*0.70702*
		ACA[5]	*95.0*	*0.05*	*103*	*0.0014*	*0.70613*	*0.70611*
426-35	D	AMA	2.9	2.08	75	0.0804	0.70737	0.70656
		ACA	95.0	0.07	67	0.0029	0.70628	0.70625
426-33	*D*	*AMA*	*2.3*	*12.7*	*342*	*0.1088*	*0.70854*	*0.70733*
		ACA	*94.8*	*0.07*	*101*	*0.0020*	*0.70589*	*0.70587*
		AMA[5]	*2.9*	*3.90*	*191*	*0.0598*	*0.70811*	*0.70744*
		ACA[5]	*95.8*	*0.07*	*102*	*0.0020*	*0.70585*	*0.70583*
426-31	D	AMA	3.5	1.84	64	0.0833	–	–
		ACA	94.9	0.45	44	0.0295	0.70614	0.70584
426-29	D	AMA	2.4	9.58	46	0.6115	0.71123	0.70478
		ACA	88.7	0.17	27	0.0184	0.70605	0.70586
426-24	*D*	*AMA*	*2.4*	*2.79*	*142*	*0.0568*	*0.70717*	*0.70656*
		ACA	*96.9*	*0.08*	*66*	*0.0037*	*0.70602*	*0.70598*
426-20	D	AMA	3.1	2.72	126	0.0625	0.70812	0.70745
		ACA	94.5	0.06	73	0.0022	0.70615	0.70612
426-18	*D*	*AMA*	*3.1*	*2.23*	*53*	*0.1226*	–	–
		ACA	*95.3*	*0.05*	*55*	*0.0024*	*0.70577*	*0.70574*
426-16	D	AMA	2.0	35.7	105	0.9865	0.72295	0.71210
		ACA	81.6	0.35	56	0.0178	0.70837	0.70817
58-12[4]	D	AMA	2.0	1.31	102	0.0376	0.70882	–
		ACA	97.9	0.08	42	0.0057	0.70820	–
426-14	D	AMA	3.7	15.7	76	0.5950	0.71338	0.70682
		ACA	92.3	0.16	54	0.0084	0.70834	0.70825
426-13	D	AMA	3.5	5.26	19	0.8416	–	–
		ACA	89.9	0.12	12	0.0258	0.70805	0.70776
426-12	D	AMA	2.1	24.0	107	0.6469	–	–
		ACA	58.6	0.67	64	0.0306	0.70847	0.70813
426-9	D	AMA	3.0	1.96	14	0.3998	0.70863	0.70416
		ACA	96.3	0.03	14	0.0063	0.70816	0.70809
426-8	D	AMA	4.0	0.76	43	0.0507	0.70887	0.70830
		ACA	90.3	0.03	28	0.0030	0.70754	0.70753
		AMA[5]	4.6	–	–	–	0.70883	–
		ACA[5]	89.7	–	–	–	0.70758	–
426-6	D	AMA	3.3	3.25	22	0.4210	–	–
		ACA	95.4	0.05	10	0.0155	0.70819	0.70802
426-5	L	AMA	4.8	7.19	179	0.1160	0.70762	0.70631
		ACA	90.8	0.13	196	0.0020	0.70574	0.70572
58-10	L	AMA	4.9	7.34	229	0.0951	0.70696	0.70589
		ACA	91.0	0.16	267	0.0017	0.70562	0.70560
58-8	L	AMA	5.3	5.62	151	0.1079	0.70746	0.70624
		ACA	88.3	0.14	185	0.0023	0.70571	0.70568

Table 2. (Contd.)

Sample no.[1]	Rock[2]	Carbonate phase[3]	Content of the phase (%)	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ measured	$^{87}Sr/^{86}Sr$ initial
Inzer Formation								
426-64	*L*	*AMA*	*5.9*	*2.43*	*208*	*0.0438*	*0.70621*	*0.70571*
		ACA	*92.9*	*0.11*	*173*	*0.0017*	*0.70557*	*0.70555*
58-1	L	AMA	4.1	0.36	175	0.0064	0.70623	0.70616
		ACA	94.9	0.33	115	0.0082	0.70567	0.70558
426-62	L	AMA	5.5	3.18	210	0.0434	0.70628	0.70578
		ACA	93.6	0.09	166	0.0015	0.70565	0.70563
63-6	*L*	*AMA*	*4.6*	*0.34*	*235*	*0.0041*	*0.70611*	*0.70606*
		ACA	*93.9*	*0.03*	*260*	*0.0004*	*0.70560*	*0.70560*
		AMA[5]	*5.7*	*0.32*	*170*	*0.0055*	*0.70594*	*0.70587*
		ACA[5]	*92.7*	*0.03*	*235*	*0.0004*	*0.70562*	*0.70562*
63-5	*L*	*AMA*	*3.8*	*0.21*	*197*	*0.0031*	*0.70623*	*0.70619*
		ACA	*95.4*	*0.04*	*213*	*0.0005*	*0.70567*	*0.70566*
426-58	L	AMA	6.4	1.33	241	0.0158	0.70641	0.70622
		ACA	87.3	0.09	302	0.0009	0.70556	0.70555
438-15	*L*	*AMA*	*5.2*	*1.35*	*596*	*0.0065*	*0.70555*	*0.70547*
		ACA	*94.8*	*0.15*	*542*	*0.0008*	*0.70539*	*0.70538*
438-12	*L*	*AMA*	*3.5*	*0.82*	*698*	*0.0030*	*0.70542*	*0.70538*
		ACA	*96.5*	*0.04*	*598*	*0.0002*	*0.70532*	*0.70532*
438-10	*L*	*AMA*	*4.9*	*0.07*	*564*	*0.0004*	*0.70546*	*0.70546*
		ACA	*94.1*	*0.02*	*517*	*0.0001*	*0.70530*	*0.70530*
438-9	*L*	*AMA*	*4.8*	*0.71*	*498*	*0.0040*	*0.70542*	*0.70537*
		ACA	*94.9*	*0.03*	*522*	*0.0002*	*0.70525*	*0.70525*
438-6	*L*	*AMA*	*5.4*	*0.61*	*618*	*0.0022*	*0.70536*	*0.70533*
		ACA	*94.6*	*0.03*	*506*	*0.0002*	*0.70533*	*0.70533*
		AMA[5]	*4.8*	*0.35*	*616*	*0.0017*	*0.70539*	*0.70537*
		ACA[5]	*95.9*	*0.36*	*512*	*0.0021*	*0.70533*	*0.70531*
438-3	*L*	*AMA*	*3.7*	*1.09*	*650*	*0.0047*	*0.70547*	*0.70541*
		ACA	*96.0*	*0.04*	*491*	*0.0002*	*0.70526*	*0.70526*
373-16	L	AMA	5.7	2.28	124	0.0538	0.70742	0.70676
		ACA	91.3	0.12	153	0.0020	0.70648	0.70646
373-13	*L*	*AMA*	*6.2*	*0.94*	*536*	*0.0052*	*0.70631*	*0.70625*
		ACA	*90.8*	*0.04*	*572*	*0.0002*	*0.70579*	*0.70579*
		AMA[5]	*5.0*	*1.90*	*589*	*0.0094*	*0.70623*	*0.70612*
		ACA[5]	*91.6*	*0.06*	*613*	*0.0003*	*0.70580*	*0.70580*
373-12	*L*	*AMA*	*5.2*	*1.12*	*387*	*0.0084*	*0.70635*	*0.70624*
		ACA	*94.5*	*0.04*	*418*	*0.0002*	*0.70565*	*0.70565*
55-4[4]	L	AMA	5.1	0.55	110	0.0145	0.70884	–
		ACA	94.9	0.06	76	0.0022	0.70888	–
441-24	L	AMA	4.6	2.39	346	0.0201	0.70579	0.70554
		ACA	94.2	0.06	249	0.0007	0.70536	0.70535
		AMA[5]	4.3	2.57	411	0.0183	0.70554	0.70536
		ACA[5]	94.9	0.04	250	0.0005	0.70539	0.70538
373-10	*L*	*AMA*	*5.5*	*1.35*	*428*	*0.0091*	*0.70621*	*0.70610*
		ACA	*94.2*	*0.04*	*439*	*0.0002*	*0.70571*	*0.70571*
441-20	*L*	*AMA*	*4.3*	*3.16*	*532*	*0.0171*	*0.70556*	*0.70535*
		ACA	*95.4*	*0.05*	*418*	*0.0004*	*0.70528*	*0.70528*

Table 2. (Contd.)

Sample no.[1]	Rock[2]	Carbonate phase[3]	Content of the phase (%)	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ measured	$^{87}Sr/^{86}Sr$ initial
441-18	*L*	*AMA*	*5.1*	*2.17*	*406*	*0.0154*	*0.70554*	*0.70535*
		ACA	*94.7*	*0.04*	*411*	*0.0003*	*0.70534*	*0.70534*
441-17	L	AMA	5.4	4.09	345	0.0347	0.70640	0.70596
		ACA	93.1	0.29	338	0.0026	0.70538	0.70536
441-15	L	AMA	4.8	10.2	572	0.0516	0.70658	0.70592
		ACA	85.6	0.18	506	0.0010	0.70541	0.70540
		AMA[5]	3.5	10.8	608	0.0511	0.70658	0.70591
		ACA[5]	87.3	0.19	503	0.0010	0.70541	0.70540
Katav Formation								
441-13	L	AMA	5.4	7.99	168	0.1393	0.70996	0.70823
		ACA	84.4	0.21	151	0.0041	0.70632	0.70627
373-9	L	AMA	5.9	7.29	213	0.1002	0.71185	0.71060
		ACA	81.2	0.19	245	0.0023	0.70756	0.70753
		AMA[5]	5.1	7.88	266	0.0868	0.71200	0.71092
		ACA[5]	81.6	0.19	239	0.0023	0.70755	0.70752
63-2[4]	L	AMA	8.6	0.47	688	0.0020	0.71199	–
		ACA	91	0.08	286	0.0008	0.71243	–
441-9	L	AMA	4.3	15.9	199	0.2338	0.71247	0.70956
		ACA	82.2	0.39	190	0.0060	0.70651	0.70644
373-5	L	AMA	4.9	13.9	214	0.1893	0.71245	0.71010
		ACA	82.1	0.07	216	0.0010	0.70821	0.70820
63-1[4]	L	AMA	8.5	0.60	142	0.0124	0.72019	–
		ACA	90.3	0.08	268	0.0008	0.72044	–
441-5	L	AMA	4.3	15.3	162	0.2766	0.71441	0.71097
		ACA	81.0	0.28	123	0.0067	0.70682	0.70674
		AMA[5]	5.8	10.8	125	0.2530	0.71415	0.71100
		ACA[5]	79.7	0.49	125	0.0115	0.70676	0.70662
55-3[4]	L	AMA	9.2	1.33	111	0.0351	0.72148	–
		ACA	89.9	0.09	160	0.0016	0.72186	–
		AMA[5]	10.0	0.88	115	0.0224	0.72141	–
		ACA[5]	88.7	0.05	159	0.0009	0.72189	–
373-3	L	AMA	5.1	18.3	176	0.3052	0.71374	0.70995
		ACA	76.3	0.34	161	0.0063	0.70845	0.70837
441-2	L	AMA	6.2	5.35	124	0.1264	0.71253	0.71096
		ACA	83.2	0.24	157	0.0045	0.70714	0.70708
		AMA[5]	5.3	11.2	144	0.2278	0.71319	0.71036
		ACA[5]	86.3	0.19	155	0.0036	0.70718	0.70714
55-2[4]	L	AMA	7.6	0.41	55	0.0220	0.72272	–
		ACA	91.2	0.24	64	0.0110	0.72472	–
373-1	L	AMA	4.7	22.0	154	0.4184	0.71824	0.71304
		ACA	71.6	0.53	115	0.0135	0.70996	0.70979
		AMA[5]	5.4	17.6	136	0.3790	0.71767	0.71296
		ACA[5]	70.1	0.52	118	0.0129	0.70991	0.70975

Note: (1) italicized are data for samples satisfying the accepted geochemical criteria (Mg/Ca ≤ 0.024, Mn/Sr ≤ 0.2, and Fe/Sr ≤ 5 for limestones and Mg/Ca ≥ 0.608, Mn/Sr ≤ 1.2, Fe/Sr ≤ 3.0 for dolostones); (2) limestone (L) or dolostone (D); (3) carbonate material dissolved in 1 N NH_4OAc (AMA phase) and residue after leaching dissolved subsequently in 10% CH_3COOH (ACA phase); (4) samples of veined calcite or dolomite; (5) data for samples independently treated in 1 N NH_4OAc.

analyzed material is not specified when necessary. Calculating the $^{87}Sr/^{86}Sr$ initial ratios, we used the following age values: 870 Ma for the Katav Formation, 835 Ma for the Inzer Formation, 780 Ma for the Min'yar Formation, and 650 Ma for the Uk formation. In the case of the Inzer and Min'yar formation, the values correspond to Pb–Pb dates obtained for their rocks. Two other values average out the age intervals adopted for corresponding formations. We did not estimate the $^{87}Sr/^{86}Sr$ initial ratios for veined calcite and dolomite, because the formation time of veins is unknown.

ISOTOPIC-GEOCHEMICAL CHARACTERIZATION OF THE KARATAU GROUP CARBONATES

Katav Formation. Clayey limestones of this formation are composed of calcite with a high percentage of fine siliciclastic admixture (10 to 24%) that consists of illite, chlorite, clastic quartz, and feldspars (Table 1). The carbonate fraction of rocks is rich in Mn and Fe, average concentrations of which correspond to 206 and 3200 ppm, respectively. In contrast, the Sr concentration is rather low, equal to 170 ppm in average and varying from 120 to 240 ppm. Calcite from fine veinlets is almost barren of Mg (<0.1%), being enriched in Mn and depleted in Fe as compared to matrix carbonates. The Sr content in this calcite from the lower part of the formation is slightly lower than in the upper one (67–155 versus 255–345 ppm). The $^{87}Sr/^{86}Sr$ initial ratios calculated for the Katav clayey limestones are high in general, and their values are higher in the Kulmas section (0.70753–0.70979) than in the Min'yar section (0.70627–0.70714). This distinction is likely related to post-sedimentation history of the rocks, which exhibit a high degree of recrystallization, stylolitization, and fracturing in the Kulmas section. Characteristic of both sections is a trend of decreasing-upward $^{87}Sr/^{86}Sr$ ratios from 0.70714 to 0.70627 in the Min'yar section and from 0.70979 to 0.70753 in the Kulmas section. The same trend is typical of veined calcite. The $^{87}Sr/^{86}Sr$ ratios in this mineral range from 0.72472 near the formation base to 0.72044–0.72189 in the middle and then to 0.71243 near its top (Table 2).

The high percentage of clay admixture in the Katav limestones can be a factor responsible for enrichment of their carbonate fraction in Mn, Fe, Rb, and radiogenic ^{87}Sr, because dia- and catagenetic transformations of clay minerals deliberate, as is known, some chemical elements from their structures (Anfimov, 1997). In fact, rather high concentrations of Mn (0.02 to 0.12%) and Fe (0.20 to 0.36%) are characteristic of pore waters in non-consolidated carbonate sediments containing 10 to 15% of clay material (Savin and Epstein, 1970). Nevertheless, $^{87}Sr/^{86}Sr$ ratios decrease upward in the Katav succession, i.e., away from underlying sandstones and shales of the Zilmerdak Formation, and according to this trend the above factor acted in combination with the other one. It is likely that

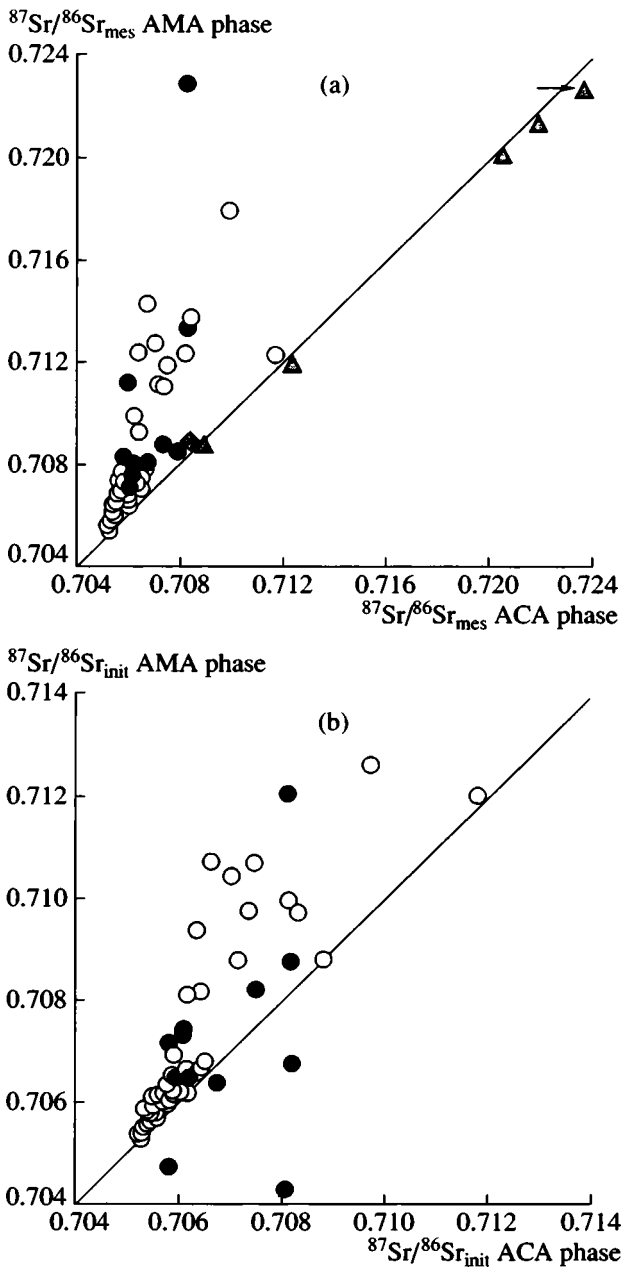

Fig. 4. Correlation diagrams for measured (a) and initial (b) $^{87}Sr/^{86}Sr$ ratios in AMA and ACA phases of carbonates from the Karatau Group (symbols as in Fig. 3).

recrystallization of the Katav limestones and enrichment of their carbonate material in trace elements and radiogenic ^{87}Sr developed also under influence of epigenetic fluids rising from underlying siliciclastic rocks during their burial and at the formation time of fracture systems in the Karatavian succession (Kuznetsov, 1998; Kuznetsov *et al.*, 2002).

Inzer Formation. All limestones of this subdivision, except for two samples (441-24 and 58-1) of dolomitized rocks, are composed of low-Mg calcite, in which

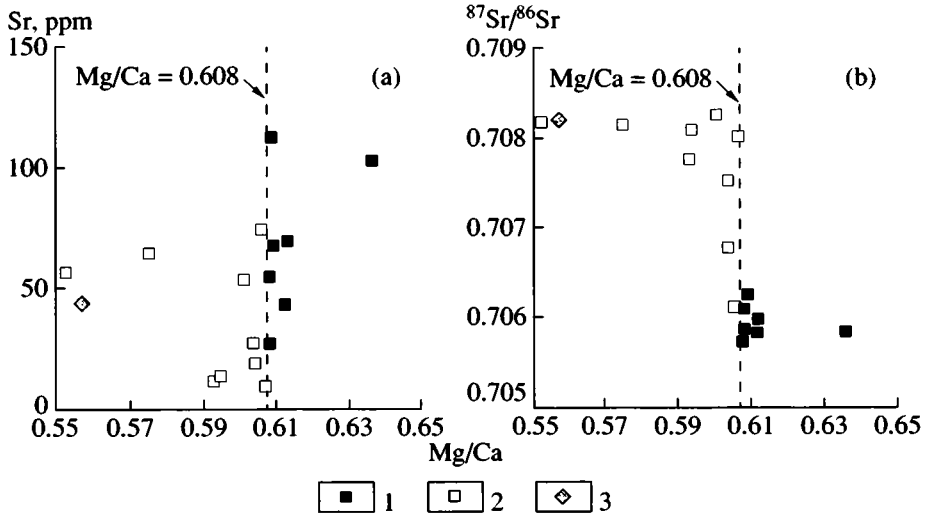

Fig. 6. Sr concentrations (a) and isotopic parameters (b) in dolostones of the Min'yar Formation plotted versus Mg/Ca ratios in the rocks: (1) high-Mg dolostones (Mg/Ca ≤ 0.608); (2) low-Mg dolostones (Mg/Ca ≤ 0.608); (3) veined dolomite.

beds in the upper one (11 to 14%). Concentration ranges of Mn and Fe are wide. Minimum concentrations of both elements are detected in pure micrites and microsparites (Mn = 22–66 ppm, Fe = 130–470 ppm). In limestones, which are enriched in siliciclastic components or dolomitized, their concentrations are greater (80–150 and 775–1010 ppm, respectively). The maximum concentration values (Mn = 102–603 ppm, Fe = 1120–3450 ppm) are established in carbonate packages of the lower subformation, which are sandwiched between siliciclastic strata, and in basal sandy limestones of the upper subformation (Figs. 5e and 5f). The Sr concentration range of 145–715 ppm characterizes a majority of Uk limestones composed of micrite and microsparite, while lower concentrations (105 to 140 ppm) are measured in limestone varieties containing the late crystalline calcite. An exception is sample 3573 from the top interval of the Kulmas section, in which the Sr concentration is 90 ppm only, although it is composed of micrite and microsparite. Dolostone and dolomitic limestone (samples 442-24 and 442-27) crowning the Shubino section contain 1.5–9.5% Mg, 110–120 ppm Mn, and 220–1120 ppm Fe. The lowermost Sr concentration (12–53 ppm) is a distinctive feature of these rocks.

The $^{87}Sr/^{86}Sr$ initial ratios range in the Uk limestones from 0.70538 to 0.70800. The minimum values (0.70538–0.70598) are characteristic of micrite to microsparite limestones with low content of siliciclastic components, whereas maximum values are typical of rocks enriched in the latter or bearing poikilitic calcite. A positive correlation of Sr isotopic parameters with Mn/Sr and Fe/Sr ratios, on the one hand, and negative correlation between $\delta^{18}O$ values and two latter ratios, on the other (Figs. 5e and 5f), are established for rocks of the last type (Podkovyrov *et al.*, 1998). This is an evidence of rock alteration under influence of meteoric fluids during the pre-Vendian break in sedimentation. The high $^{87}Sr/^{86}Sr$ ratios (0.71050–0.71187) in dolomitic limestones sampled near the formation top suggest that these rocks have been recrystallized under influence of fluids enriched in radiogenic ^{87}Sr.

Source of epigenetic ^{87}Sr in the Karatau Group carbonates. Taking into consideration the Karatau Group structure, we assume that epigenetic fluids enriched in radiogenic ^{87}Sr migrated in its carbonate horizons along zones of tectonic decompaction from basal sandy to clayey deposits of the Zilmerdak Formation and from the Katav succession of clayey-carbonate sediments at the time of their burial and diagenesis. This assumption is based on the following facts. (1) Experiencing transformation after burial of siliciclastic and clayey-carbonate deposits, clay minerals are able to deliberate considerable amounts of Mn, Fe, Rb, and radiogenic ^{87}Sr, which contaminate the associated carbonate sediments. (2) The $^{87}Sr/^{86}Sr$ ratios in the suited calcite veins decrease upward in the Katav succession, i.e., away from the Zilmerdak Formation top, and directly point to the source of fluids enriched in radiogenic ^{87}Sr. (3) The $^{87}Sr/^{86}Sr$ ratio ranges in easily soluble phases leached from clay minerals of the Zilmerdak and Inzer shales correspond to 0.7220–0.7225 and 0.7126–0.7181, respectively (Gorokhov *et al.*, 1995; Kuznetsov *et al.*, 2002). (4) The $^{87}Sr/^{86}Sr$ ratios measured in late calcite and dolomite from veins in the Karatau Group rocks range from 0.70820 to 0.72472. (5) The $^{87}Sr/^{86}Sr$ maximum ratios have been detected in surface AMA phases of limestone (0.71292) and dolostone (0.72295) samples from the Karatavian succession. On the other hand, it is also possible that the Karatavian carbonates became additionally enriched in radiogenic ^{87}Sr after interaction with meteoric and ground waters that leached this component from nearby siliciclastic deposits. Influence of this factor is most obvious in terminal

horizons of the Min'yar and Uk formations, especially in the last case.

POSTSEDIMENTARY TRANSFORMATIONS IN THE KARATAU GROUP CARBONATES AND NEW DATA ON $^{87}Sr/^{86}Sr$ RATIOS IN THE LATE RIPHEAN SEAWATER

The primary Mn/Sr and Fe/Sr ratios in marine carbonate sediments are low. Chemistry and isotopic composition of sediments change however at the dia- and epigenetic stages, when they interact with elision, ground, and meteoric waters, which are enriched, as compared to seawater, in Mn, Fe, and radiogenic ^{87}Sr leached from associated siliciclastic deposits (Goldberg, 1963; Blatt *et al.*, 1980; Kharaka and Thordsen, 1992; Stueber *et al.*, 1993; Chaudhuri and Clauer, 1993). In addition, postsedimentary chemical transformations in carbonate sediments are controlled by values of water-rock distribution coefficients of trace elements relative to Ca. This coefficient is <1 for Sr, but >1 for Mn and Fe (Pingitore, 1978; Kretz, 1982; Banner and Hanson, 1990; Vahrenkamp and Swart, 1990). As a result, carbonate sediments became enriched in Mn and Fe after low-temperature recrystallization, but partially depleted in Sr, the isotopic composition of which also changes therewith (Brand and Veizer, 1980; Yudovich *et al.*, 1980; Gorokhov, 1996). Consequently, the absolute values of Mn/Sr and Fe/Sr ratios proper and dependences between these parameters, on the one hand, and $^{87}Sr/^{86}Sr$ ratio, on the other hand, are indicative of the secondary alteration degree in Precambrian carbonates (Veizer *et al.*, 1983; Derry *et al.*, 1992; Gorokhov *et al.*, 1995; Melezhik *et al.*, 2001; Semikhatov *et al.*, 2002). As it is demonstrated in practice, a combination of petrographic, lithologic, and geochemical methods is the best way to elucidate secondary alterations in carbonates and to establish the threshold values of geochemical criteria, which enable selection of best preserved samples appropriate for reconstruction of seawater $^{87}Sr/^{86}Sr$ in the past.

The Rb–Sr systematics of carbonate rocks from type and reference Riphean sections studied in the Urals and Siberia (Kuznetsov *et al.*, 1997; Semikhatov *et al.*, 1998, 2002) allows us to suggest the following critical parameters for selection of the least altered limestones: $Mn/Sr \leq 0.20$, $Fe/Sr \leq 5.0$, and $Mg/Ca \leq 0.024$. Although these parameters can be not universal and applicable in all possible cases of Rb–Sr system transformation in carbonates, they are the strictest among all others suggested so far in publications. In our opinion, they actually ensure a strict selection of the least altered ("best") limestones.

Many researchers who studied secular $^{87}Sr/^{86}Sr$ variations on paleooceans believe that dolostones represent not reliable source of information (Derry *et al.*, 1989, 1992; Asmerom *et al.*, 1991; Kaufman *et al.*, 1993). It is proved now, however, that the early dolomitization of calcareous sediments may develop under seawater influence in shallow-water basins (Ricketts, 1983; Vahrenkamp and Swart, 1990; Land, 1992; Montanez and Read, 1992b; and others), and that these dolostones, if they are lacking signs of intense alteration, are able to retain the Sr isotope signature of sedimentation medium (Kupecz and Land, 1991; Montanez and Read, 1992a; Gorokhov *et al.*, 1995, 1998). Nevertheless, inadequate data on behavior of trace elements during lithogenesis are a serious obstacle for establishing criteria indicative of the least altered dolostones. Data obtained for carbonate rocks of the Karatau Group offer a progress in this direction.

For instance, the studied samples (441-24 and 58-1) from the Inzer Formation imply that a weak early-diagenetic dolomitization did not change the primary Sr isotope composition in the rocks. Despite an elevated Mg content (1.1–1.4%) exceeding the range of 0.1–0.8% characteristic of other rocks from the stratigraphic interval under consideration, the $^{87}Sr/^{86}Sr$ ratios in dolomitic varieties do not differ from those in associated "pure" limestones (Table 2). In addition, dolomicrites relicts present in some early-diagenetic cherty concretions from high-Mg dolostones of the Min'yar Formation suggest that dolomitization of calcareous sediments took place immediately after their deposition, and that this process was likely under control of fluids, which were close in chemical and isotopic composition to sedimentation medium of primary carbonate material. Thus, dolostones of this origin can retain the Sr isotopic signature of seawater. It is also important that the high-Mg dolostones of the Min'yar Formation show, in distinction from low-Mg varieties, the better-preserved sedimentary structures, a lower degree of recrystallization, and a higher Sr content. Accordingly, the above data suggest that the high-Mg dolostones, which originated in the course of the early-diagenetic dolomitization, are lacking serious secondary alterations and preserve primary geochemical characteristics. In contrast, large zonal crystals present in low-Mg dolostones of the Min'yar Formation and low Mg and Sr concentrations typical of these rocks mean that they experienced a partial dedolomitization. In other words, the chemical composition of dolostones is indicative of the preservation extent of their Rb–Sr systems. Critical values of $Mg/Ca = 0.608$, $Mn/Sr = 1.2$, and $Fe/Sr = 3.0$ discriminate the high-Mg dolostones from low-Mg varieties (Table 1), and interaction of carbonate rocks with epigenetic fluids added, as is shown above, the radiogenic ^{87}Sr to their composition. Consequently, dolostones with lower Mg/Ca, but higher Mn/Sr and Fe/Sr ratios can be regarded as significantly altered and inappropriate for inference of seawater Sr isotope composition. It is remarkable as well that the above value of critical Mn/Sr ratio is close to that (1.0 to 1.5) formerly accepted for the geochemical criterion discriminating the least- and stronger-altered samples of Upper Proterozoic limestones (Asmerom *et al.*, 1991; Derry *et al.*, 1992; Kaufman *et al.*, 1993).

Thus, we establish the following threshold values of geochemical criteria applicable for selection of those analyzed samples, which can retain the seawater Sr isotopic signature: Mn/Sr ≤ 0.20, Fe/Sr ≤ 5.0, and Mg/Ca ≤ 0.024 for limestones, and Mn/Sr ≤ 1.2, Fe/Sr ≤ 3.0, and Mg/Ca ≥ 0.608 for dolostones. Analytical results for samples satisfying these criteria are italicized in Tables 1 and 2.

Clayey limestones of the Katav Formation, which are the lowermost carbonate rocks in the Karatavian succession, show a wide range of Sr isotopic parameters and high Mn/Sr and Fe/Sr ratios far-exceeding their threshold values. Accordingly, the Rb–Sr systems of these rocks were reset in the course of postsedimentary alterations, and the corresponding analytical results (Table 2, Fig. 7) are omitted from further consideration as inappropriate for reconstruction of seawater Sr isotopic composition. In distinction, the majority of samples from the Inzer Formation (16 of 22) satisfy the accepted geochemical criteria. The $^{87}Sr/^{86}Sr$ initial ratios in these 16 samples are within the range of 0.70555–0.70566 in both studied sections. Different ranges of this parameter have been established for the "best" lower Inzer limestones from the Min'yar and Kulmas sections (0.70525–0.70538 and 0.70565–0.70580, respectively; Fig. 7). The higher ratios in rocks of the latter section seem to reflect a more intense recrystallization of corresponding limestones, which bear veins of late calcite with a high $^{87}Sr/^{86}Sr$ ratio (0.70888). Formally, however, these rocks satisfying criteria ensuring their geochemical retentivity are left as legal for subsequent analysis. Epigenetic alterations are more intense in 25 studied samples from the Min'yar Formation (17 of dolostones and 8 of limestones). As a result, only four samples of high-Mg dolostones appeared to be fitting the accepted geochemical criteria. The $^{87}Sr/^{86}Sr$ initial ratios, which have been obtained for these samples and represent a series of values increasing upward from 0.70574 to 0.70611, are consistent in general with the upper limit of this parameter in seawater that has been estimated (Fig. 7) based on data for altered limestones sampled near the formation base (< 0.70560) and top (<0.70600). It is likely that the interval of 0.70574–0.70611 adequately characterizes the Sr isotopic composition in seawater of the pre-Shubino interval of the Min'yar time. Rocks of the Uk Formation, which crown the carbonate succession of the Karatau Group and have been deposited in the western limb of the Bashkirian meganticlinorium before the pre-Vendian break in sedimentation, are intensely recrystallized under influence of ground and meteoric waters. Among 15 studied samples from this formation, only three fit criteria of their geochemical retentivity. The $^{87}Sr/^{86}Sr$ initial ratios in these limestone samples range from 0.70538 to 0.70580. Thus, the Karatau Group is informative with respect to the Sr isotopic composition in seawater only for a limited time span of the Late Riphean (pre-Shubino interval). Constraints in this case are as follows: (1) basal horizons of the group are composed of the Zilmerdak siliciclastic sediments and of the Katav clayey limestones, which are badly altered; (2) between the Min'yar and Uk Formations there is a hiatus presumably spanning a considerable stratigraphic interval; and (3) the Uk Formation is separated from Vendian basal strata by the terrigenous Krivaya Luka Formation and subsequent hiatus.

$^{87}Sr/^{86}Sr$ RATIO VARIATIONS IN THE LATE RIPHEAN OCEAN AND GEODYNAMIC BACKGROUND

The Lakhanda Group of the Uchur-Maya region and its analogues in the Turukhansk area, which have been deposited in the early Karatavian time (Gorokhov *et al.*, 1995; Semikhatov *et al.*, 1998, 2002), represent, together with the Uralian Inzer, Min'yar, and Uk formations spanning a considerable interval of the Late Karatavian time (Gorokhov *et al.*, 1996; Kuznetsov *et al.*, 1997; Kuznetsov, 1997, and this work), the key subdivisions that enable reconstruction of $^{87}Sr/^{86}Sr$ secular variations in the Late Riphean seawater. The prime importance of these subdivisions is defined by the fact that they represent predominantly carbonate successions, which have been only slightly altered at the epigenetic stage. In addition, they are rather well calibrated in terms of chronometry, have been studied using a uniform procedure that included the two-stage chemical treatment of samples, which satisfy strict criteria of their geochemical retentivity. Unfortunately, the necessary data have been obtained so far only for three discrete time spans of the Late Riphean (1030–1000, 850–770, and 690–640 Ma) characterizing in sum about 37% of the latter. The Sr isotopic characterization of intervals between these time spans are based either on data published for carbonates from other regions, or partially on interpolation.

The Late Riphean history of seawater $^{87}Sr/^{86}Sr$ variations followed that of the Middle Riphean, when fluctuations of this parameter were fairly quiet (see Semikhatov *et al.*, 2002, for review and references). According to available data, the seawater $^{87}Sr/^{86}Sr$ ratio in the early Middle Riphean, 1280–1270 Ma ago, was comparable with that of the Early Riphean time and did not exceed 0.7046–0.7047. Later on, this ratio increased up to 0.7049 1260 Ma ago and then up to 0.7052–0.7056 1200 Ma ago. That increase was a response to events of the Elsevirian accretionary orogeny, which was of a subglobal significance and commenced about 1300–1250 Ma ago (McLelland *et al.*, 1996; Rivers, 1997; Rivers and Corrigan, 2000). The seawater Sr isotope composition in the period of 1200–1050 Ma is not established yet, although the Shawinigan (1190–1140 Ma) and Ottawan (1080–1020 Ma) collisional impulses of that period (Rivers and Corrigan, 2000; Carr *et al.*, 2000) suggest a further increase of $^{87}Sr/^{86}Sr$ ratio in the ocean. The Middle Riphean maximum of this ratio (0.70592) is recorded in the Malgina Formation of the Uchur-Maya region (1050–1040 Ma ago,

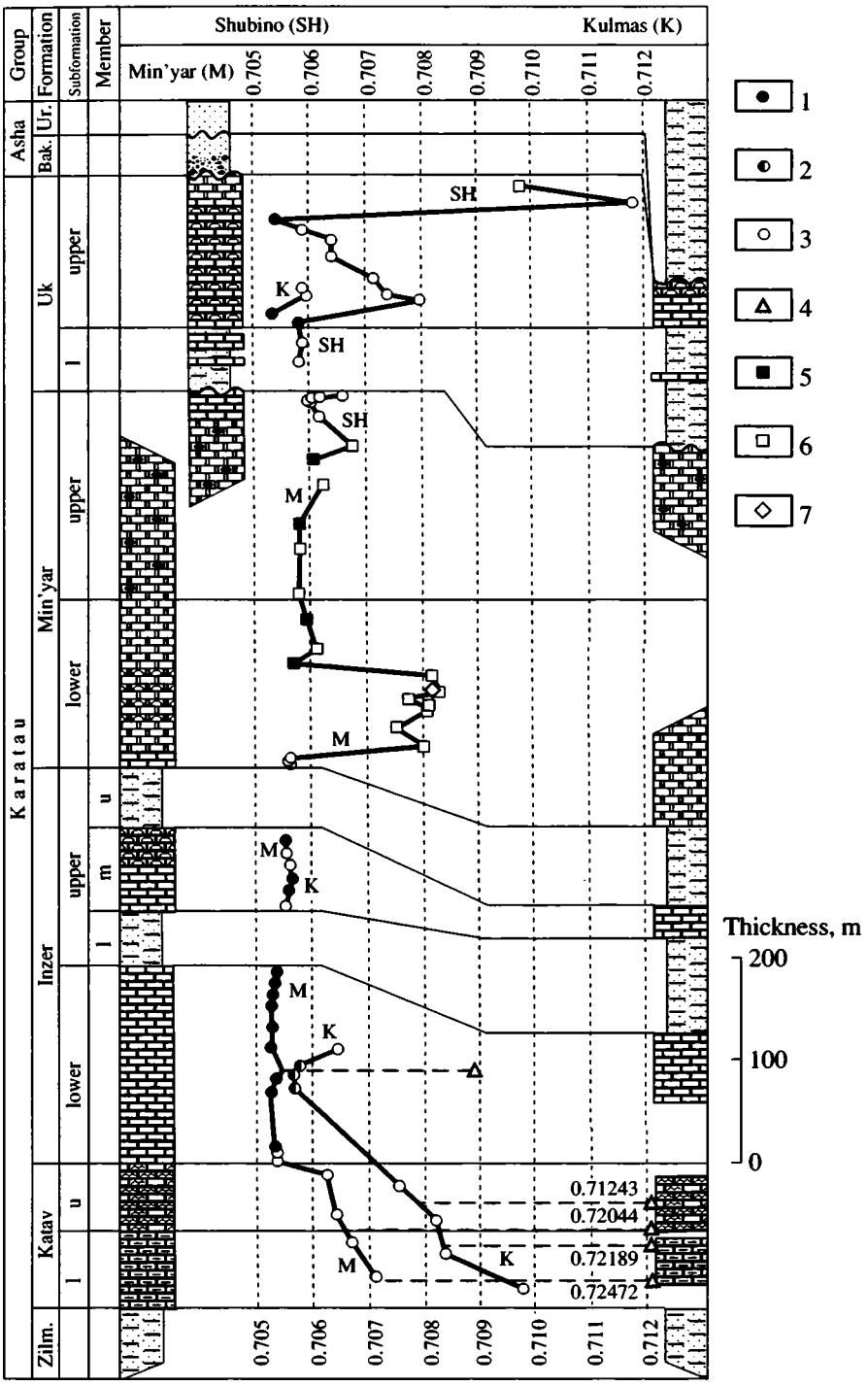

Fig. 7. $^{87}Sr/^{86}Sr$ variations in carbonates of the Karatau Group: (1) limestones satisfying the accepted geochemical criteria (Mn/Sr ≤ 0.2, Fe/Sr ≤ 5.0, Mg/Ca ≤ 0.024); (2) limestones satisfying the accepted geochemical criteria but containing late calcite generation; (3) limestones non-satisfying the accepted geochemical criteria; (4) veined calcite; (5) dolostones satisfying the accepted geochemical criteria (Mn/Sr ≤ 1.2, Fe/Sr ≤ 5.0, Mg/Ca > 0.608); (6) dolostones non-satisfying the accepted geochemical criteria; (7) veined dolomite. Horizontal dotted lines denote sampling levels of veined calcite with indicated $^{87}Sr/^{86}Sr$ ratios outside the scale limit; (K) Kulmas, (M) Min'yar, (SH) Shubino sections; other abbreviations and symbols as in Fig. 2.

Semikhatov *et al.*, 2002). By the end of the Middle Riphean (1030 Ma ago), the ratio somewhat decreased down to 0.70567–0.70585 (Sukhaya Tunguska Formation of the Turukhansk region, Gorokhov *et al.*, 1995). Afterward, at the very beginning of the Late Riphean (1030–1000 Ma), this tendency continued, as it is evi-

dent from data on carbonates from the Lakhanda Group of the Uchur-Maya region and its Turukhansk analogues (Gorokhov *et al.*, 1995; Semikhatov *et al.*, 1998, 2002; Bartley *et al.*, 2001). In the "best" samples from above subdivisions, the $^{87}Sr/^{86}Sr$ ratio ranges from 0.70519 to 0.70569, decreasing in general to 0.70523–0.70527 in sediments deposited about 1000 Ma ago. It is also remarkable that transition from the ascending Middle Riphean trend to the descending one of the Early Karatavian time took place at the time of the Ottawan (1080–1020 Ma) and Rigolet (1000–980 Ma) collision events of the Grenville orogeny, which resulted in a global-scale collision and in the emergence of the Rodinia supercontinent (Rivers, 1997; Carr *et al.*, 2000; Ketchum and Davidson, 2000; Martignole *et al.*, 2000).

Explanation to the paradoxical seawater $^{87}Sr/^{86}Sr$ ratio decrease at the time of the Grenville orogeny termination and initial erosion of orogens created by that time was found after analysis of corresponding data of geodynamics, geochemistry, and isotopic geochronology (Semikhatov *et al.*, 2002). Factors responsible for the paradox are as follows. (1) pre-Grenvillian mantle rocks were apparently widespread in the Grenville orogenic structures (Restrepo-Pace *et al.*, 1997; Wareham *et al.*, 1998; Geraldes *et al.*, 2000; Van Schmus, 2000; Dickin, 2000; Carr *et al.*, 2000). (2) A considerable mantle Sr flux into the World Ocean is admissible in connection with the opening and evolution of the Paleo-Asian ocean and its branches at the early Late Riphean time (V. Khain *et al.*, 1997; Khain *et al.*, 1999, 2002, 2003; Khomentovskii and Postnikov, 2001; Khudoley *et al.*, 2001; Fedotova and Khain, 2002) and with concurrent development of the Gaios Ocean within the future Gondwana (1200–800 Ma, Strieder and Suita, 1999) and global Monrovia Ocean around Rodinia (Weil *et al.*, 1998). A high juvenile Sr influx into seawater is inferable also from the Sm-Nd systematics of the Middle-Upper Riphean boundary deposits (Li and McCulloch, 1996; Podkovyrov *et al.*, 2002) and from $\mu_1 = {}^{238}U/{}^{204}Pb$ calculated for the concurrent carbonate sediments based on the one-stage model of Pb isotope evolution (Ovchinnikova *et al.*, 2002). (3) Peak events of metamorphism are asynchronous in diverse Grenville structures (Andersson *et al.*, 1999; Bruguier *et al.*, 1999; Carr *et al.*, 2000; Martignole *et al.*, 2000) and, what is most important, the erosion rate of the latter was low and not constant. Besides the short-term post-metamorphic peaks, that rate was more than one order of magnitude lower than the upwarping rate in Alps and Himalayas during the last 40 Ma (Cosca *et al.*, 1998; Ketchum and Davidson, 2000; Martignole *et al.*, 2000; Rivers and Corrigan, 2000), when the $^{87}Sr/^{86}Sr$ ratio sharply increased in the Late Tertiary oceans mainly under influence of this leading factor. (4) In addition to peneplanation of tectonic relief in the eastern Laurentia about 980–920 Ma ago (Cosca *et al.*, 1992; Carr *et al.*, 2000), a vast transgression of the early Karatavian time (Semikhatov and Raaben, 1994, 1996) enlarged the carbonate platform areas favorable for expansion of benthic cyanobacterial ecosystems and thus reduced the continental runoff. (5) As is commonly accepted, phases of an abrupt sea-level rise are interrelated with activation of sea-floor spreading, and contribution of the latter to seawater Sr isotopic parameters could be also perceptible. The development of intense crustal magmatism ($\varepsilon_{Nd}(t)$ from –2.4 to –11 ± 2) in the West Congolese belt 1000–920 Ma ago (Tack *et al.*, 2001) or collisional events 880–860 Ma ago in the Yenisei Ridge, intrusions of crustal granitoids in Alaska and Taimyr, where they are 950 to 800 Ma old (Vernikovsky and Vernikovskaya, 2001; Vernikovsky *et al.*, 2002; Vernikovskaya *et al.*, 2002), and incipient subduction that developed in the Goianides Ocean 900–850 Ma ago (Brito Nevis *et al.*, 1999) were unable to compensate the effect of factors listed above.

A segment of seawater $^{87}Sr/^{86}Sr$ variation curve that corresponds to the Lakhanda Group time span is the best substantiated one for the Proterozoic (Semikhatov *et al.*, 1998, 2002; Bartley, 2001). For the subsequent period of about 150 Ma long, there are only fragmentary data on the Sr isotopic record. First, these are the $^{87}Sr/^{86}Sr$ initial ratios 0.70570–0.70622 established for altered dolostones of the Shorikha and Turukhansk formations in the terminal part of the Riphean succession of the Turukhansk region and ranging in age approximately from 950 to 900 Ma (Gorokhov *et al.*, 1995). Next, there are Sr isotopic data on limestones and dolostones from members I_4 to I_{10} of the Atar Group in western Africa (Veizer *et al.*, 1983). Carbonates from these members were intensely recrystallized at the stages of their burial and meteoric diagenesis (Fairchild *et al.*, 1990), as it follows from their highly variable and elevated Mn/Sr, Fe/Sr, and $^{87}Sr/^{86}Sr$ ratios. The only sample from the base of Member I_6 is acceptable for determination of that ratio upper limit in seawater according to minimal values of Mn/Sr = 0.43 and Fe/Sr = 3.3 measured in it. The Rb–Sr dates obtained for the clay fraction < 2 μm from members I_6 and I_5 are 874 ± 23 and 890 ± 37 Ma, respectively (Clauer, 1981). These age determinations are approximate, however, because the studied fractions contain non-cogenetic illite generations (Gorokhov *et al.*, 1994, 2001). In this work, the age of 870 Ma consistent with data on stromatolites from a lower part of the Atar Group (Bertrand-Sarfati, 1972; Knoll and Semikhatov, 1998) is conventionally adopted for Member I_6. Using this value in calculations, we obtain the $^{87}Sr/^{86}Sr$ initial ratio of 0.70558 to characterize limestones of Member I_6.

Data on the Inzer Formation of the southern Urals and Bitter Springs Formation of central Australia (Figs. 8 and 9) define termination of the gap in the Sr isotopic record under consideration. The $^{87}Sr/^{86}Sr$ initial ratios calculated for the "best" limestones from the lower part of the lower Inzer Subformation (850–840 Ma) range from 0.70525 to 0.70534. The $^{87}Sr/^{86}Sr$ ratios measured in the whole-rock samples of dolostones and sulfates from lower member of the Bitter Springs For-

mation correspond to 0.70592–0.70611 and 0.70569–0.70610, respectively, and quickly grow upward reflecting either the basin isolation, or epigenetic alteration of samples under influence of meteoric waters (Hill and Walter, 2000; Walter *et al.*, 2000). The Bitter Springs Formation is defined to be 840–830 Ma old based on U–Pb zircon dates (827 ± 6 and 849 ± 9 Ma) for basic dikes comagmatic with volcanic rocks of the formation (Zhao *et al.*, 1994; Wingate *et al.*, 1998; Hill and Walter, 2000). As the values under discussion are obtained for different materials (ACA phases of "best" limestones in one case and whole-rock samples of dolostones and sulfates in the other), we may speak only about their general similarity, bearing in mind that data from the Urals define the $^{87}Sr/^{86}Sr$ ratio in the Inzer Ocean, whereas data from Australia determine only the upper limit of that ratio in seawater. Materials considered above imply that the $^{87}Sr/^{86}Sr$ ratios in seawater of the terminal Lakhanda (1000 Ma) and initial Inzer (850 Ma) times were almost identical (0.70523–0.70527 and 0.70525–0.70534), whereas 950–900 to 870 Ma ago they were below the values of 0.70622 and 0.70558, respectively.

Persistently low, insignificantly changeable $^{87}Sr/^{86}Sr$ ratios in seawater of the early Karatavian time were likely controlled by a high mantle Sr influx into the World Ocean that prevailed over the continental influx almost throughout 200 m.y. Facts in favor of this inference are as follows. (1) Widespread eruptions of plateau basalts with $^{87}Sr/^{86}Sr$ ratios from 0.7043 to 0.7050 were characteristic of the West Congolese belt 1000–920 Ma ago (Tack *et al.*, 2001). (2) Eruptions of basalts with positive $\varepsilon_{Nd}(t)$ values were related to the first stage of the Adamastor-Brasilide Ocean opening between 950 and 800 Ma ago (Dalziel, 1997; Pedrosa-Soares *et al.*, 2001). (3) Evolution of the Gaios (Strieder and Suita, 1999) and Paleo-Asian oceans (Kovalenko *et al.*, 1999; Khain *et al.*, 2002, 2003; Fedotova and Khain, 2002) was in progress at that time. (4) An island-arc system and ophiolites with low $^{87}Sr/^{86}Sr$ initial ratios (0.7022–0.7030) and positive $\varepsilon_{Nd}(t)$ values appeared at the site of the future Nubian Shield 890–800 Ma ago (Reischmann, 2000). (5) The Goianides Ocean opened 900–850 Ma ago in South America (Brito Nevis *et al.*, 1999) concurrently with origination of ophiolites in Mauritania and Anti-Atlas. (6) An intense rifting and/or intrusion of basaltic dike swarms 890 to 820 Ma old were typical of many large regions (Wingate *et al.*, 1998; Li *et al.*, 1999; Karlstrom *et al.*, 2000; Yarmolyuk and Kovalenko, 2001; Pedrosa-Soares *et al.*, 2001; Sklyarov *et al.*, 2001). (7) Passive margins of the Siberian craton experienced transformation that was followed 850–800 Ma ago by development of extended island arcs along periphery of the Paleo-Asian Ocean (Khain *et al.*, 2003; Fedotova and Khain, 2002). (8) Ophiolite belts of the Central Asian foldbelt originated 800 Ma ago (Kovalenko *et al.*, 1999; Khain *et al.*, 2003). (9) Volcanic island arcs emerged in the Paleo-Asian Ocean at the beginning of the Late Riphean (Kovalenko *et al.*, 1999; Fedotova and Khain, 2002; Khain *et al.*, 2003). (10) An extensive transgression of the early Karatavian time decelerated erosion of Grenville orogens and reduced influx of continental material into the World Ocean. All these facts evidence not a short-term "mantle event at 900 Ma ago" (Veizer *et al.*, 1983) but a series of large-scale events of the early Karatavian time, which favored influx of low-radiogenic Sr into the ocean (Figs. 8 and 9). Regional accretionary-collisional events and intrusions of post-collisional granitoids (880 to 870 Ma old) (Nozhkin *et al.*, 1999; Vernikovskaya *et al.*, 2002) along the Siberian craton western edge were unable to affect the global situation.

Returning to analysis of data from the Urals, we would like to stress that the Sr isotopic parameters of carbonates from the Inzer and Min'yar formations point to an important change in the character of seawater Sr-isotope variations about 830 Ma ago. The former "quiet," maybe nearly horizontal segment of the variation curve changed into its ascending branch, the origin of which corresponds to boundary horizons between the lower and upper Inzer subformations and termination is at the Shubino Member base. During the corresponding time span (830–775 Ma), the seawater $^{87}Sr/^{86}Sr$ ratio rose from 0.70555 to 0.70611, and then, at the Shubino time (ca. 770 Ma), it dropped again below 0.70600 (Fig. 9).

Reconstruction of seawater Sr-isotope variations within the subsequent time spans of the Late Riphean is complicated by an ambiguous interpretation of age relationships between three main sources of relevant information—the Shaler Group of Arctic Canada, Akademikerbreen Group of Spitsbergen, and Karatau Group of the southern Urals (Melezhik *et al.*, 2001).

The Shaler Group includes the Glenelg Formation (1400 m) of carbonate-siliciclastic composition and the Reynolds Point (760 m), Minto Inlet (210 m), Wanniatt (590 m), and Kilian (550 m) formations of predominantly carbonate composition. Deposits of the group discordantly overlie basic volcanics bearing baddeleyite that yielded the U–Pb age value of 1267 ± 2 Ma, whereas baddeleyite from dikes crosscutting the group is 723 ± 3 Ma old (Heaman *et al.*, 1990; Rainbird *et al.*, 1994). To precise the maximum age limit, a lower part of the Shaler Group was correlated first with the Ray Group of Arctic Canada, where the U–Pb age of 1080 Ma was obtained for clastic zircons, and next, taking into account palinspastic reconstructions, with an upper part of the Kollanna Group in Australia, where the U–Pb age of a tuff horizon is defined to be 802 ± 10 Ma (Rainbird *et al.*, 1994). Data of C-isotope chemostratigraphy seem to be more informative in this respect. They show that $\delta^{13}C$ values from +3 to +5‰ PDB, which are typical of the Reynolds Point Formation, are followed by higher values of +6 to +8‰ PDB of the overlying Minto Inlet and Wanniatt formations (Asmerom *et al.*, 1991; Kaufman and Knoll, 1995).

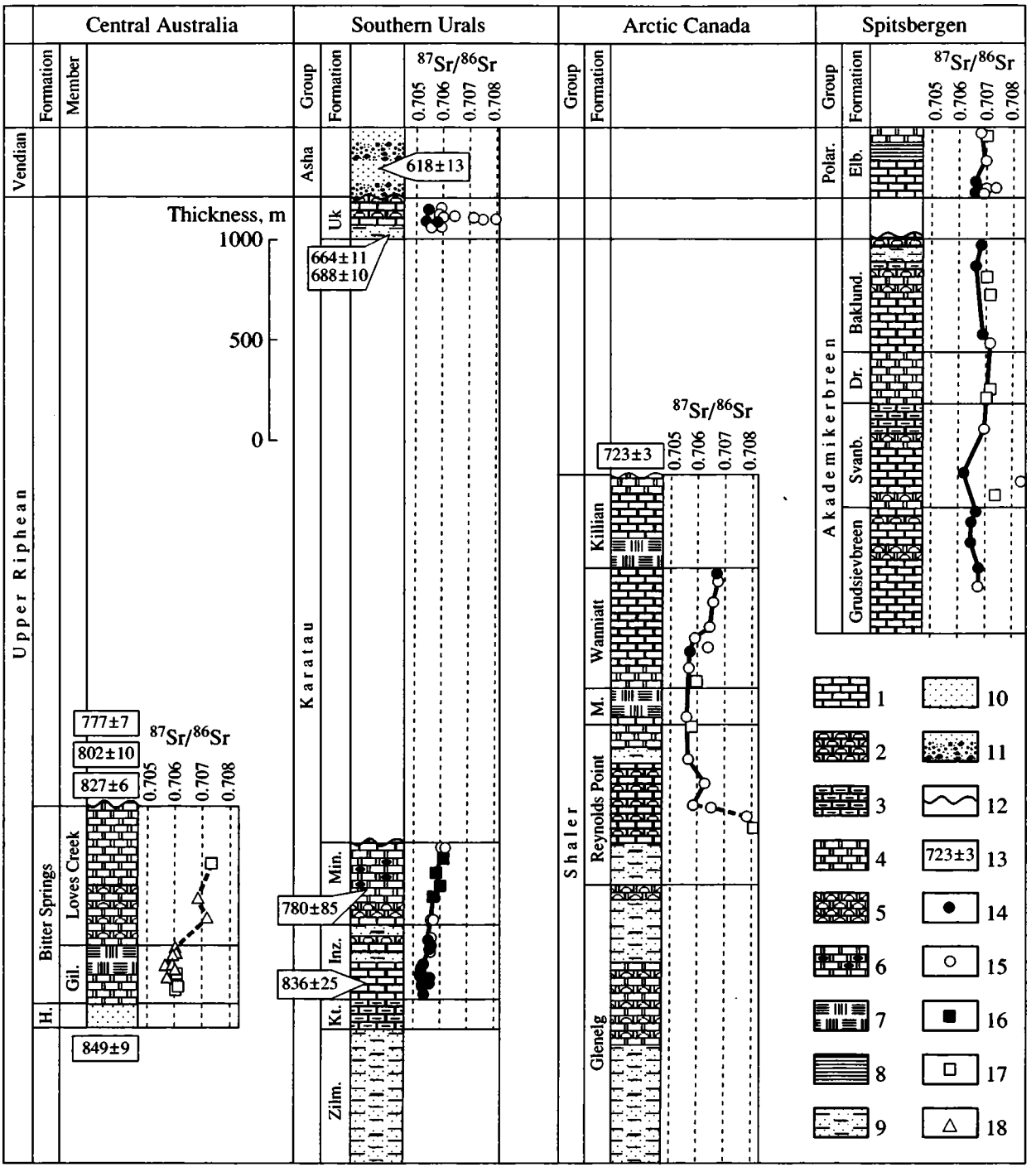

Fig. 8. Sr-chemostratigraphy and age characterization of key sections of the upper Upper Riphean and Lower Vendian. Lithology: (1) limestone; (2) stromatolitic limestone; (3) clayey limestone; (4) dolostone; (5) stromatolitic dolostone; (6) dolostone with cherty concretions; (7) sulfates (evaporites); (8) tillites; (9) sandy-clayey deposits; (10) sandstone; (11) sandstone with gravelstone interlayers; (12) stratigraphic unconformity; (13) isotopic age (see explanations in the text). Samples in terms of geochemical criteria accepted in this work: (14, 15) limestones satisfying (14) and non-satisfying (15) the criteria; (16, 17) dolostones satisfying (16) and non-satisfying (17) the criteria; (18) sulfates. Abbreviations: (Baklund.) Baklundtoppen; (Gil.) Gillen; (Dr.) Draken; (Zilm.) Zilmerdak; (Inz.) Inzer; (Kt.) Katav; (M.) Minto Inlet; (Min.) Min'yar; (Polar.) Polarisbreen; (Svanb.) Svanbergfjellet; (H.) Heavitree; (Elb.) Elbobreen.

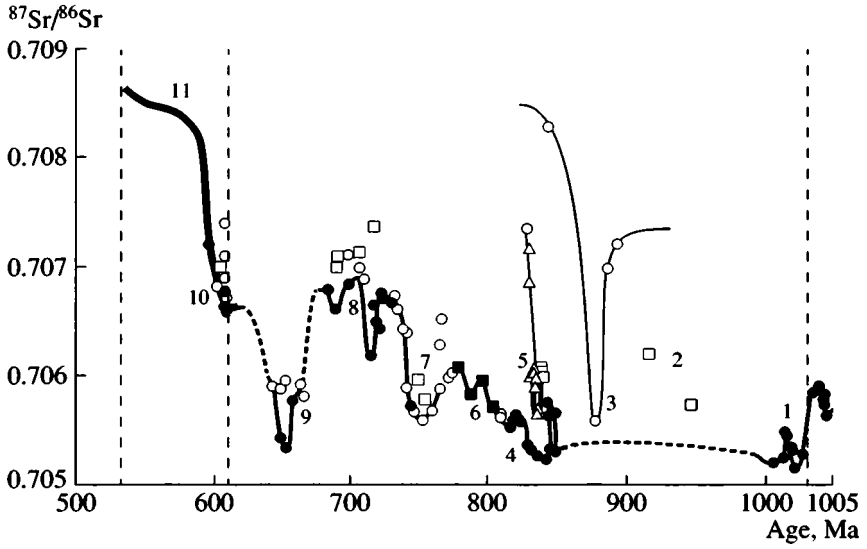

Fig. 9. Standard curve of $^{87}Sr/^{86}Sr$ in the Late Riphean seawater: (1) Kerpyl and Lakhanda groups and their analogues in the Turukhansk region (Gorokhov *et al.*, 1995; Semikhatov *et al.*, 1998, 2002; Bartley *et al.*, 2001); (2) Shorikha and Turukhansk formations of the Turukhansk region (Gorokhov *et al.*, 1995; Bartley *et al.*); (3) Member I_6 of the Atar Group (Veizer *et al.*, 1983); (4) Inzer Formation of the Karatau group (this work; Kuznetsov *et al.*, 1997); (5) Bitter Springs Formation (Walter *et al.*, 2000; Hill and Walter, 2000); (6) Min'yar Formation of the Karatau Group (this work); (7) Shaler Group (Asmerom *et al.*, 1991); (8) Akademikerbreen Group (Derry *et al.*, 1989; 1992); (9) Uk Formation (this work); (10) Polarisbreen Group (Kaufman *et al.*, 1993); (11) Nama and Vitvley groups (Kaufman *et al.*, 1993); symbols as in Fig. 8.

When evaluated in the global context (Jacobsen and Kaufman, 1999; Kah *et al.*, 1999; Bartley *et al.*, 2001), these values suggest that the Reynolds Point Formation is younger than 850–800 Ma. On the other hand, carbonates of the Min'yar Formation yield $\delta^{13}C$ values mostly ranging from –2.2 to +2.7‰ PDB (Podkovyrov *et al.*, 1998), though its terminal horizons correspond to an ascending trend of this parameter with a maximum value of +5.9‰ PDB at the top of the Shubino Member. This trend seems to be implying that carbonate sediments of the Reynolds Point, Minto Inlet, and Wanniatt formations were deposited in the post-Min'yar time 770–730 Ma ago. It is reasonable to remind therewith that deposits of this age are unknown in the southern Urals.

Data on the Sr isotope composition in the Shaler Group are known for 17 limestone and dolostone samples from three middle formations of this subdivision (Fig. 8). The Mn/Sr and $^{87}Sr/^{86}Sr$ ratios widely range in these samples from 0.12 to 9.3 and from 0.70561 to 0.70876, respectively (Asmerom *et al.*, 1991). Only two limestone samples from the Wanniatt Formation satisfy criteria of geochemical retentivity accepted in this work. The $^{87}Sr/^{86}Sr$ initial ratios calculated for the bulk carbonate material of this samples are equal to 0.70575 and 0.70674. However, a minimum value of this ratio (0.70561) is established for an altered limestone sample (Mn/Sr = 0.67) from the lower part of the Minto Inlet Formation (Fig. 8). If, despite the altered status of the rock, the last value reflects nevertheless the Sr isotope composition in sedimentation medium, then we may compare this and above values with data on carbonates from the Min'yar Formation, which immediately predate the indicated carbonates of the Shaler Group. Such a comparison shows that the seawater $^{87}Sr/^{86}Sr$ ratios decreased gradually from 0.70600 to 0.70561 since the end of the Min'yar time during the subsequent 10–15 m.y., and afterward, at the deposition time of the upper Wanniatt Formation, they became growing again and attained the level of 0.70674 at about 730 Ma ago (Fig. 9).

The next interval of the standard curve of seawater $^{87}Sr/^{86}Sr$ variations can be elucidated based on data for the Akademikerbreen Group of Spitsbergen (Derry *et al.*, 1989, 1992). The successive Grusdievbreen (650 m), Svanbergfjellet (500 m), Draken (250 m), and Baklundtoppen (500 m) carbonate formations of the group conformably rest on the terrigenous Veteranen Group, and, according to recent data (Fairchild and Hambrey, 1995; Kaufman *et al.*, 1993, 1997; Harland, 1997), their eroded top is overlain with a hiatus by the Polarisbreen Group. Near the base of the latter, there are tillites of the Varangerian Glaciohorizon that is 600–590 Ma old in opinion of many researchers (Knoll and Walter, 1992; Knoll, 2000; Semikhatov, 2000, and references therein). To our regret, isotopic dates for rocks of the Akademikerbreen Group have not been obtained. In the 1970s, the group was correlated, based on microphytoliths and stromatolites, with the Riphean stratotype of the Urals (Raaben and Zabrodin, 1972; Raaben, 1975), but the results obtained need to be verified. As is established (Semikhatov, 1974; Khomentovskii, 1976; Yakshin, 1985), stratigraphic significance of microphytoliths used for correlation is negligi-

ble, and stromatolites of the group are represented mainly by endemics and taxa identified in an open nomenclature. In the 1980s, some horizons of the group were dated based on silicified microfossils, but these results need in a verification as well. Thus, the location of the Akademikerbreen Group in stratigraphic scale strongly depends on chemostratigraphy.

Postsedimentary alterations in carbonate rocks of the Akademikerbreen Group are insignificant (Derry *et al.*, 1989, 1992). Among 12 limestone and 5 dolostone samples, which have been studied, 8 limestone samples are satisfying our criteria of geochemical retentivity. The $^{87}Sr/^{86}Sr$ initial ratios range in these limestone samples mostly from 0.70646 to 0.70686 and decrease down to 0.70620 in the lower part of the Svanbergfjellet Formation only (Fig. 8). The $^{87}Sr/^{86}Sr$ ratios (0.70646–0.70677) and $\delta^{13}C$ values (+5.9‰ PDB) in middle carbonate strata of the Grusdievbreen Formation are practically identical to these parameters (0.70662–0.70674 and +5.0 to +6.0‰ PDB) characterizing a terminal part of the Wanniatt Formation of the Shaler Group (Derry *et al.*, 1989; Asmerom *et al.*, 1991). This concordance between chemostratigraphic characteristics suggests that the above formations are correlative, and that the middle Grusdievbreen strata were deposited about 730 Ma ago immediately after accumulation of the Wanniatt Formation. The exact isotopic age of terminal horizons of the Akademikerbreen Group is unknown. This group is separated from the basal carbonate Member E1 of the Polarisbreen Group by a hiatus, the time span of which is unknown (Fairchild and Hambrey, 1995; Harland, 1997). Chemostratigraphic data on the Karatau Group suggest that accumulation of the Akademikerbreen Group terminated ca. 690 Ma ago, at the time of initial deposition of the Uralian Uk Formation. Actually, carbonates of the Baklundtoppen Formation crowning the Akademikerbreen Group are characterized by $^{87}Sr/^{86}Sr$ initial ratios of 0.70662 to 0.70686 and by $\delta^{13}C$ values of +4.2 to +8.4‰ PDB, whereas these parameters in the lower Uk carbonates deposited 690–660 Ma ago are much lower: 0.70582 to 0.70595 and +1.4 to +2.6‰ PDB, respectively. In the upper Uk carbonates with the age range of 660–640 Ma, the same parameters decrease further and are within the intervals of values from 0.70538 to 0.70580 (Fig. 9, Table 2) and from +0.7 to +2.5‰ PDB (Podkovyrov *et al.*, 1998).

Consequently, the Uk Formation and upper horizons of the Akademikerbreen Group cannot be regarded as age equivalents because of a considerable distinction between their Sr- and C-isotope parameters. On the other hand, the former also differs in these parameters from the carbonates of Member E1 at the base of the Polarisbreen Group, characteristic of which are the $^{87}Sr/^{86}Sr$ initial ratios equal to 0.70661–0.70683 and predominantly high (+5 ± 1‰) $\delta^{13}C$ values (Knoll *et al.*, 1986; Kaufman *et al.*, 1993; Fairchild and Spiro, 1987). The negative $\delta^{13}C$ values have not been established in Member E1 underlying the lower tillites of the Varangerian Glaciohorizon of Spitsbergen, but they are typical of Member 19 below the same tillites of Greenland, and this discrepancy has been regarded as evidence in favor of a hiatus below tillites of Spitsbergen (Fairchild and Hambrey, 1995; p. 223). It is likely therefore that sediments of the Uk Formation were deposited in the period of that hiatus between the Akademikerbreen and Polarisbreen groups, which was presumable as long as 60–70 m.y. Nevertheless, it is impossible to suggest a confident Sr-isotopic characterization of corresponding time span, because the lower Uk interval is predominantly composed of siliciclastic deposits intercalated with carbonates, which have been significantly altered at the epigenetic stage (Fig. 7). The corresponding time span that was presumably as long as 30 m.y. separates the "best" upper carbonates of the Akademikerbreen Group from those of the upper Uk Subformation.

Turning now to consideration of geodynamic and other factors, which controlled the $^{87}Sr/^{86}Sr$ ratio variations in seawater of the Late Karatavian time, we should remind that a distinct ascending trend of that ratio is recorded in "best" carbonates samples from the upper Inzer and Min'yar strata deposited 830–775 Ma ago. It is logical to assume that this trend is interrelated with a series of accretionary and collisional events of that time. These events and associated granitoid magmatism are detectable in many regions. A collision of microcontinent with island arcs or continent happened about 850–800 Ma ago in the northwestern Alaska-Taimyr region (Vernikovsky and Vernikovskaya, 2001). The Tarim massif basement originated 829 ± 60 Ma ago in connection with a collision between two microcontinents (Guo *et al.*, 2001). In the Tuva-Mongolia belt, continental blocks collided 800 Ma ago either with each other and/or with the Siberian craton margin, and these events were accompanied by development of an island-arc system in oceanic peripheral zone (Salnikova *et al.*, 2001; Sklyarov *et al.*, 2001; Fedotova and Khain, 2002). Events of collision, accretion, and local subduction 830–800 Ma ago characterize an important stage in evolution of the Paleo-Asian Ocean (Khain *et al.*, 2002, 2003). Syntectonic granitoids of the Baikal-Muya fold belt originated 800 Ma ago in connection with a powerful collision that terminated the Early Baikalian tectogenesis in the southeast of Siberia (Rytsk *et al.*, 2001). In the Baikal-Vitim branch of that ocean, a microcontinent subducted under Siberian craton and island arc fragments became docked with a (micro)continent 850 Ma ago (Khomentovskii and Postnikov, 2001), while in South America the Gaios Ocean closure is dated back to 800 Ma (Strieder and Suita, 1999). The mantle flux of low-radiogenic Sr, intrusions of mafic dikes with a mantle isotopic signature, and expansion of carbonate platforms along passive continental margins counterbalanced the continual growth of seawater $^{87}Sr/^{86}Sr$ ratio in the period from 830 to 770 Ma. Nevertheless, the net Sr influx into the World Ocean was enriched in radiogenic isotope at that time.

In response to global geodynamic changes around 770 Ma ago, the ascending trend of seawater $^{87}Sr/^{86}Sr$ ratio turned for a short but significant period (from ca. 775 to 740 Ma) into the descending one. The latter is recorded in rocks of the Shubino Member (less than 0.70600) and in lower horizons of the Shaler Group, in which the ratio decreases from 0.70611 down to 0.70561–0.70575. A subglobal extension that commenced at that time (Powell *et al.*, 1993; Hoffman, 1991; Karlstrom *et al.*, 2000; Prave, 1999; Brito Nevis *et al.*, 1999) triggered a series of the following important events. (1) East Gondwana (Australia, India and Antarctic) split off from Laurentia 750 Ma ago, and this resulted in emergence of the Paleo-Pacific Ocean and in a later (700–720 Ma ago) intense rifting in South America (Powell *et al.*, 1993; Colporn *et al.*, 2002; Brito Nevis *et al.*, 1999). (2) In the western (present-day coordinates) margin of Laurentia, a high-scale development of extended rift structures up to 2500 km long was associated 760–720 (or 780–720) Ma ago with activation of basaltic magmatism that was very intense in places and with emergence of basins floored by oceanic crust (Ross, 1995; Hoffman, 1991; Colporn *et al.*, 2002). (3) Volcanic island arcs and ophiolites appeared in the future Arabia-Nubian Shield, East Egyptian desert, and Taimyr correspondingly 770–736, about 780, and 750–720 Ma ago (Reischmann, 2000; Loizenbauer *et al.*, 2001; Vernikovsky and Vernikovskaya, 2001; Vernikovsky *et al.*, 2002). (4) A volcanic arc of the Andean type (750 ± 2.5 Ma old) emerged along the eastern margin of Gondwana (Torsvik *et al.*, 2001). (5) Partition of East Gondwana, South China continental block, and Laurentia took place at about 700 Ma ago (Li *et al.*, 1996). (6) Rifting and associated intrusions of basic dikes and sills were characteristic 770–720 Ma ago of the South China platform, Tuva, South Africa, and North Canada (Heaman *et al.*, 1990; Hoffman *et al.*, 1996; Li *et al.*, 1999; Yarmolyuk and Kovalenko, 2001). It should be mentioned that Asmerom *et al.* (1991) who rightfully connected the Shaler event with an intense development of juvenile crust and large-scale oceanic rifting attributed all phenomena to the period of 840–810 Ma ago. According to recent results (Karlstrom *et al.*, 2000), these events took place later, however, within the time span of 762–723 Ma. Concurrent events of opposite significance can be exemplified by collision and post-collisional granitoids of the Yenisei Ridge, which are 750–720 Ma old, by the microcontinent-island arc collision 740 Ma ago in the north of the Mozambique belt, and by the intra-plate granitoids of eastern Egypt, which are 780 Ma old (Reischmann, 2000; Loizenbauer *et al.*, 2001; Vernikovsky *et al.*, 2002; Vernikovskaya *et al.*, 2002). Contribution of these events to the total substance balance in World Ocean likely was negligible.

Somewhat higher seawater $^{87}Sr/^{86}Sr$ ratios (0.70646–0.70686) characterize the next period of the Late Riphean history (from 740 to 690 Ma), i.e., the accumulation period of upper horizons in the Wanniatt Formation of the Shaler Group and of a greater portion of the Akademikerbreen Group (boundary horizons of the Grusdievbreen and Svanbergfjelet formations excluded). These elevated ratios are in a harmony with the following geodynamic events of that time. In the indicated period, or more precisely between 736 and 715 Ma, oceanic basins of the future Arabia-Nubian Shield closed, and, after convergence of India and East Africa 740–720 Ma ago, first the Mozambique and then the Farusian oceans disappeared (Reischmann, 2000; Torsvik *et al.*, 2001; Brito Nevis *et al.*, 1999). In the Paleo-Asian Ocean, it was a time of disappearance of large island arcs and back-arc basins and of a concurrent (735–715 Ma) intensification of supra-subduction magmatism (Sklyarov *et al.*, 2001; Fedotova and Khain, 2002; Khain *et al.*, 2003). In some regions, these led to intrusion of crustal granites and to replacement of basaltic volcanism by calc-alkaline magma eruptions (Kovalenko *et al.*, 1999; Reischmann, 2000; Loizenbauer *et al.*, 2001; Kuzmichev, 2001; Fedotova and Khain, 2002). In particular, extended ensialic island arcs with abundant eruptions of andesitic, dacitic, and rhyolitic magmas originated along the periphery of Paleo-Asian Ocean 730–700 Ma ago (Kuzmichev *et al.*, 2001; Khain *et al.*, 2003). Less significant events of the same type are exemplified by accretion of island arcs around continental blocks in the Urals-Mongolian belt and by a microcontinent collision with the Siberian craton 760–720 Ma ago, or by post-collisional intrusions of crustal granites in the Yenisei Ridge, which are 730 to 718 Ma old according to results of U–Pb dating (Vernikovsky *et al.*, 2002; Vernikovskaya *et al.*, 2002; Sklyarov *et al.*, 2001; Khain *et al.*, 2003).

The paleoclimatic factor could be also responsible for the seawater $^{87}Sr/^{86}Sr$ ratio growth during the Late Karatavian time. According to paleomagnetic data, the Rodinia fragments migrated across a set of latitudes during the Late Karatavian time. That migration must control the annual precipitation and intensity of weathering and erosion in continental provenances, in other words, the continental runoff value. A tight interrelation between the seawater $^{87}Sr/^{86}Sr$ ratios and climate humidity is convincingly demonstrated for the Late Permian and Triassic (Korte *et al.*, 2003). To our regret, the idea of an increased continental runoff during the Late Karatavian time is inconsistent with data implying that abundance of carbonate stromatolitic formations was at that time twice greater than in the Early Karatavian time (Raaben and Semikhatov, unpublished data). It is logical to think that these data suggest a lateral expansion of carbonate sedimentation and concurrent reduction of settings with siliciclastic deposits. Unfortunately, the comparative data on corresponding terrigenous formations are unknown.

It is more difficult to explain a negative excursion of $^{87}Sr/^{86}Sr$ ratios from 0.70646 to 0.70620 that is established based on three samples from boundary horizons of the Grusdievbreen and Svanbergfjelet formations and happened, as inferred above, about 720–710 Ma

ago (Figs. 8 and 9). If the inferred age is correct, the excursion is concurrent to the Sturtian glaciation that must unavoidably result in a large glacioeustatic sea-level drop (as large as 100 m according to assessment of N.M. Chumakov). An immediate consequence would be intensification of continental runoff and a considerable growth rather than decline of seawater $^{87}Sr/^{86}Sr$ ratios.

If the situation under discussion is not an artefact, one may assume, to explain it, that the mantle flux of low-radiogenic Sr was extraordinary high 720–700 Ma ago owing to global extension events, and that exactly these events controlled the Sr isotopic balance in the World Ocean during 20 m.y. However, the evaluated intensity of events implies that such an assumption is highly improbable. In order to solve the posed problem, it is necessary to specify age relationships between that excursion and Sturtian glaciation, to seek for marks of glacioeustatic sea-level drop in the Akademikerbreen Group succession, and to elucidate stratigraphic relations between these marks and $^{87}Sr/^{86}Sr$ ratios in "best" carbonates of the group. Work of Fairchild and Hambrey (1995) devoted to the overlying Polarisbreen Group and Varangerian glaciation is a good example of such an analysis.

The terminal segment of the proposed standard curve of $^{87}Sr/^{86}Sr$ ratio variations in the Late Riphean ocean is mainly based on data obtained for the Uk Formation of southern Urals. This segment characterizes a quick decline of that ratio within the time span of 690–640 Ma from the range of 0.70662–0.70682 established for upper horizons of the Baklundtoppen Formation to values ranging from 0.70585 to 0.70592 in altered limestones of the lower Uk Subformation and from 0.70538 to 0.70580 in "best" limestones of the upper one. Such a considerable drop of $^{87}Sr/^{86}Sr$ ratios reflects a regime of tectonic extension in many regions, the concurrent development of new island arcs, and an intense magmatic activity, that of the mantle origin included. In such geodynamic circumstances, the Baltia split off from Laurentia about 650–630 Ma ago. This event was initiated by rifting and intrusions of tholeiitic dike swarms of mantle origin first in the western margin of Scandinavia and somewhat later in Labrador and Appalachians (Hoffman, 1991) and terminated 620–605 Ma ago with the opening of the future Iapetus Ocean (Torsvik *et al.*, 1996; Svenningsen *et al.*, 2001). In the Paleo-Asian Ocean, it was a formation time of extended volcanic arcs: a vast belt of extension and associated pyroxene-gabbro intrusions 630 to 620 Ma old, which yield $\varepsilon_{Nd}(t)$ mantle values (Fedotova and Khain, 2002), are situated now in the Transbaikal region; corresponding island-arc complexes 637 ± 5 Ma old and somewhat older (697 ± 4 Ma) granitoids with a mantle Sm-Nd signature and $^{87}Sr/^{86}Sr$ ratio of 0.70385 are established in the Yenisei ophiolite belt (Vernikovsky *et al.*, 2001). In the Nubian Shield, acid volcanic eruptions took place 670–618 Ma ago (Reischmann, 2000). Island arcs of the Tocantins belt in South America originated 640 Ma ago, partially after reworking of older continental and oceanic crust (Brito Nevis *et al.*, 1999). The Cadomian island arc that is extraordinary long (from New Scotland and Avalon to the Polar and, maybe, Central Urals) also originated within the time span under consideration, as it is evident from dates ranging around 675 Ma and characterizing basic magmatism and ophiolites of this structure (Scarrow *et al.*, 2001).

The mentioned and analogous events in other regions controlled the seawater $^{87}Sr/^{86}Sr$ ratio decrease. At the same time 660–640 Ma ago, some regional events acted in opposite direction. For instance, a gradual convergence of Gondwanan fragments and their collision 664 ± 20 Ma ago gave rise to the syncollisional calc-alkaline magmatism in the East Egyptian desert (Loizenbauer *et al.*, 2001). It was also a time of active collision in the Baikal-Vilyui branch of the Paleo-Asian Ocean (Khomentovskii and Postnikov, 2001). As is established in the Tocantins province (Pedrosa-Soares *et al.*, 2001), a series of terranes and island arcs became docked somewhat later, 630–620 Ma ago, with a continental massif just formed in the Adamastor Ocean that closed 600–570 Ma ago.

These and other large-scale events were a preamble of the global Panafrican orogeny that gave rise to the sharp unprecedented rise of $^{87}Sr/^{86}Sr$ ratios in the Vendian and Early Cambrian seawater (Kaufman *et al.*, 1993). This rise is however out of the scope of this paper.

CONCLUSIONS

(1) Carbonate sediments of Karatau Group have been insignificantly altered after deposition, especially in northwestern sections (town of Min'yar and Shubino Settlement) of the Bashkirian meganticlinorium of the southern Urals. During burial and tectonic deformations, isotopic-geochemical systems in lower limestones of the group and in low-Mg dolostones of the Min'yar Formation have been disturbed under influence of elision fluids derived from sandy-clayey sediments (from the Zilmerdak Formation rocks and from siliciclastic constituents of the Inzer Formation). In addition, carbonates of the Uk Formation and upper strata of the Min'yar Formation were partly recrystallized under influence of meteoric waters.

(2) The Sr isotope variations in the Upper Riphean carbonate rocks are studied much better than in other stratigraphic intervals of the Proterozoic. Nevertheless, there is no standard curve, which can be adopted broadly to characterize these variations in the Late Riphean seawater. This is a consequence of the following reasons: the available curves are plotted based on data obtained for discrete carbonate successions, correlation between which is not unambiguous; significance of epigenetic alterations in carbonate rocks is underestimated; criteria used to control the geochemical retentivity of studied carbonates are not uniform; and the

procedure of selective dissolution and removal of epigenetic carbonate phases have been used in exceptional cases only.

(3) The new standard curve of $^{87}Sr/^{86}Sr$ ratio variations in the Late Riphean ocean is predominantly based on analytical results obtained for the Karatau Group carbonates from the southern Urals and for the Lakhanda Group carbonates from the Uchur-Maya region and for age equivalents of the latter from the Turukhansk region. These subdivisions have a confident chronometric calibration, and their rocks are slightly altered and satisfy strict criteria of geochemical retentivity. In addition, the studied samples have been subjected to preliminary chemical treatment in order to remove the late diagenetic carbonate phases. The resultant curve of detailed seawater $^{87}Sr/^{86}Sr$ variations are plotted for three intervals of the Late Riphean ranging from 1030 to 1000, from 850 to 770, and from 660 to 640 Ma. The intermediate intervals are characterized based on critical evaluation of published Sr isotope parameters of corresponding rocks and partially on interpolation. Such an approach made it possible to present a modified version of the standard curve of $^{87}Sr/^{86}Sr$ variations in the Late Riphean seawater, which, in distinction from earlier variants, is lacking several inconsistence and former disadvantages.

(4) During the first half of the Late Riphean (1030–810 Ma) the seawater $^{87}Sr/^{86}Sr$ ratio was low, ranging from 0.70519 to 0.70566. Later on, up to about 775 Ma ago, that ratio rose to 0.70611, but afterward, during the subsequent 30 m.y., it decreased again down to 0.70561–0.70565. Within the time span of 740–690 Ma, the ratio in question ranged from 0.70646 to 0.70686 with a short-term drop down to 0.70620 about 720 Ma ago. At the end of the Late Riphean (660–640 Ma), the ratio lowered to 0.70538–0.70580 to become growing up to 0.70840–0.70860 during the Vendian and earliest Cambrian.

(5) Secular variations of Sr isotope composition in paleooceans depend not only on balance between the mantle and continental Sr flux into the World Ocean, but also on compositional changes in provenances, on the character of volcanism, and/or on changeable intensity of weathering and erosion under influence of paleoclimatic events and changes in the atmosphere composition (CO_2–O_2 balance is of prime importance). Consequently, these variations are sensitive indicators of secular variations in proportions between various microelements and nutrients brought into the World Ocean. The noticeably higher values and fluctuation amplitudes of $^{87}Sr/^{86}Sr$ ratio in seawater of the late Late Riphean, as compared to these parameters of the earlier intervals of the Proterozoic, reflect changes of a comparable scale in seawater composition. These changes influenced for sure the Precambrian biota on the eve of the Vendian and Phanerozoic. In other words, data on seawater $^{87}Sr/^{86}Sr$ variations in the World Ocean allow us to analyze secular trends of global biosphere evolution.

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Jurassic Planktonic and Benthic Cosmopolitan Foraminiferal Communities of Central Turkey

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Abstract—Jurassic sections of central Turkey (Ankara region) are described. As is established for the first time, they yield diverse assemblages of benthic and planktonic foraminifers, which are represented by 10 genera and different morphotypes. The Jurassic (Sinemurian-Pliensbachian to Oxfordian) sections are largely composed of terrigenous and carbonate rocks. In the transgressive-type section, basal conglomerates and coastal sandstones (Lower Jurassic) are overlain by relatively deep-water mudstones and tuffites (Middle Jurassic), which grade upsection into hemipelagic to pelagic radiolarian cherts and micritic limestones (Upper Jurassic-Valanginian). Remains of ammonoids occur throughout the section. All foraminiferal communities lack endemic taxa characteristic of Tethyan carbonate platforms and consist exclusively of geographically widespread and cosmopolitan species. Cosmopolitan taxa of benthic foraminiferal communities are important for interregional correlation of Jurassic sediments. Planktonic species first discovered in the Lower Jurassic interval are diverse, variable in morphology, and substantially widen our knowledge of the origination time and evolution of this important fossil group.

Key words: Jurassic, Turkey, stratigraphy, foraminifers, plankton, benthos, cosmopolitan forms, endemic forms, correlation.

INTRODUCTION

The Jurassic sediments are of a limited distribution in the Eastern Mediterranean. Their small exposures are locally known in northeastern Egypt, the Sinai Peninsula, Israel, Jordan, central and western Turkey, Kuwait. Iran, and Iraq (Kuznetsova *et al.*, 1996; Kuznetsova, 2000). The exposed sections are composed of carbonate and terrigenous sediments with subordinate cherty and siliceous-carbonate rocks.

In the Ankara area, the Jurassic-Cretaceous sedimentary sequences unconformably rest upon regionally metamorphosed rocks of the Karakaya Complex (Kocyigit, 1992). The Jurassic non-metamorphosed transgressive succession begins here with basal conglomerates grading upward into sandstones and siltstones with limestone interbeds, which are crowned with radiolarian cherts and pelagic limestones. The succession is divided into regional stratigraphic units: Lalelik Formation (Batman *et al.*, 1978), Hasonoglan and Akbaiir formations (Akyurek *et al.*, 1984), and Ankara Group (Kocyigit, 1987). Exposures of Jurassic sediments are known north and northeast of Ankara (Balum and Kosrelik areas) and in the southern part of the Ankara City (Alacaatli and Beitepe sites) (Fig. 1).

Stratigraphic subdivisions under consideration are distinguished mainly based on ammonoids, which are common and, sometimes, abundant in limestone interbeds traditionally attributed to the "Ammonitico Rosso" facies. The Upper Jurassic sediments contain radiolarians, which were studied recently (Bragin and Tekin, 1999). The radiolarian assemblages from the Ankara area are well correlative with coeval zonal assemblages of the Western Mediterranean (Baumgartner *et al.*, 1995) and thus are useful for a detailed subdivision of Upper Jurassic sequences.

In the course of study of Late Jurassic radiolarians from the Alacaatli area (Bragin and Tekin, 1999), it was established that Oxfordian sediments contain diverse foraminiferal assemblages (Kuznetsova and Bragin, 1999), which include the zonal planktonic species *Globuligerina oxfordiana* (Grigelis) in association with characteristic benthic forms. Data on foraminifers have been used for the identification of Oxfordian sediments and for their subsequent subdivision into substages and foraminiferal zones. The studied foraminiferal assemblages are of a cosmopolitan taxonomic composition and completely lack the Tethyan endemic forms. A high adaptation level and short existence time of cosmopolitan taxa from these foraminiferal assemblages are favorable for a large-scale, even subglobal, correlation of Jurassic marine sequences (Kuznetsova and Bragin, 1999).

As was shown recently (Kuznetsova *et al.*, 2002), foraminifers occur permanently, being sometimes abundant, in limestone interbeds within the Lower-Middle Jurassic portion of the studied sedimentary suc-

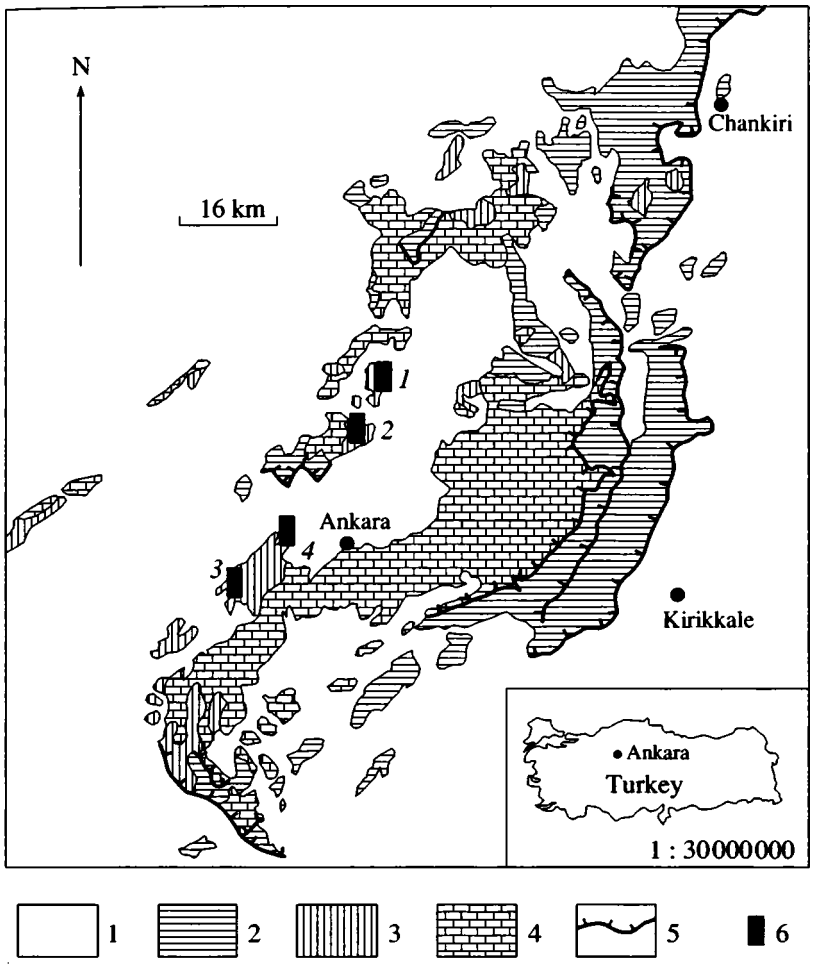


Fig. 1. Geological scheme of the Ankara area and localities of the examined Kosrelik Kizigi (*1*), Kosrelik (*2*), Alacaatli (*3*), Tahar Tepe (*4*) sections: (1) upper Campanian and younger sediments; (2) Upper Cretaceous ophiolitic melange; (3) Jurassic and Lower Cretaceous sediments; (4) Karakaya Complex (Carboniferous-Triassic); (5) thrusts; (6) section localities.

cession in the Ankara and other areas of Turkey. Their detailed study is undoubtedly of high importance for the regional stratigraphy of southern Turkey. The purpose of this work is to characterize the composition and distribution of foraminifers in Jurassic sections of the Ankara area and to define their potential for the regional stratigraphy and facies analysis.

STRATIGRAPHY, LITHOLOGY, AND SEDIMENTATION SETTINGS OF JURASSIC SEQUENCES IN THE ANKARA AREA

The Jurassic foraminifers characterize four sections of the Ankara area at the Kosrelik Kizigi, Kosrelik, Alacaatli, and Tahar Tepe sites (Figs. 2, 3). Bragin described and sampled these sections during the geological excursion organized in 2000 by Professors F. Alkaya and U.K. Tekin. The lithological description of these sections with indication of stratigraphic levels with foraminifers is given below. The age determination of the studied sediments is based on ammonoids (Prof. Alkaya, personal communication). Near the town of Balum, northeast of Ankara, the succession of Lower-Middle Jurassic sediments is approximately 300 m thick and can be subdivided into the lower detrital-carbonate (110 m), middle detrital-pyroclastic (100 m), and upper detrital-carbonate (70–80 m) parts.

The oldest, Lower Jurassic beds are exposed in the vicinity of the Kosrelik Kizigi Village. Their succession from the base upward is as follows (Fig. 2):

Bed 1: yellowish gray to brownish gray polymictic conglomerates with well-rounded pebbles (2–5 cm across) of granites, quartzite, vein quartz, metamorphosed limestones, cherts, and shales. The lower contact is unexposed. The apparent thickness is 15 m.

Bed. 2: yellowish gray medium- to fine-grained quartz-feldspar sandstones, slightly carbonate, massive to obscurely flaggy sometimes. Clastic material is well to moderately sorted, represented by angular to subangular quartz (up to 50% of the rock) and feldspar (up to 25%) grains, by fragments of mainly quartzites and

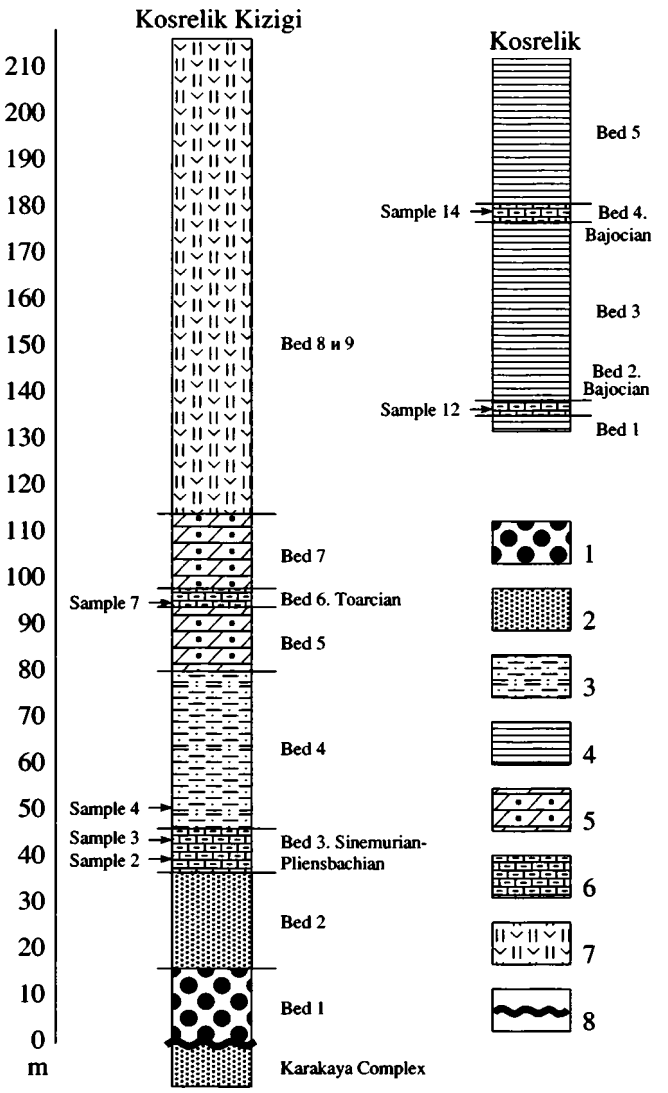


Fig. 2. Lower and Middle Jurassic sections of the Ankara area (bed numbers and their ages based on foraminifers are shown to the right of stratigraphic columns; sampling levels for foraminiferal analysis are shown to the left of the columns): (1) conglomerates; (2) sandstones; (3) siltstones; (4) mudstones; (5) sandy marls; (6) cloddy limestones of the Ammonitico Rosso type; (7) tuffaceous siltstones; (8) unconformity.

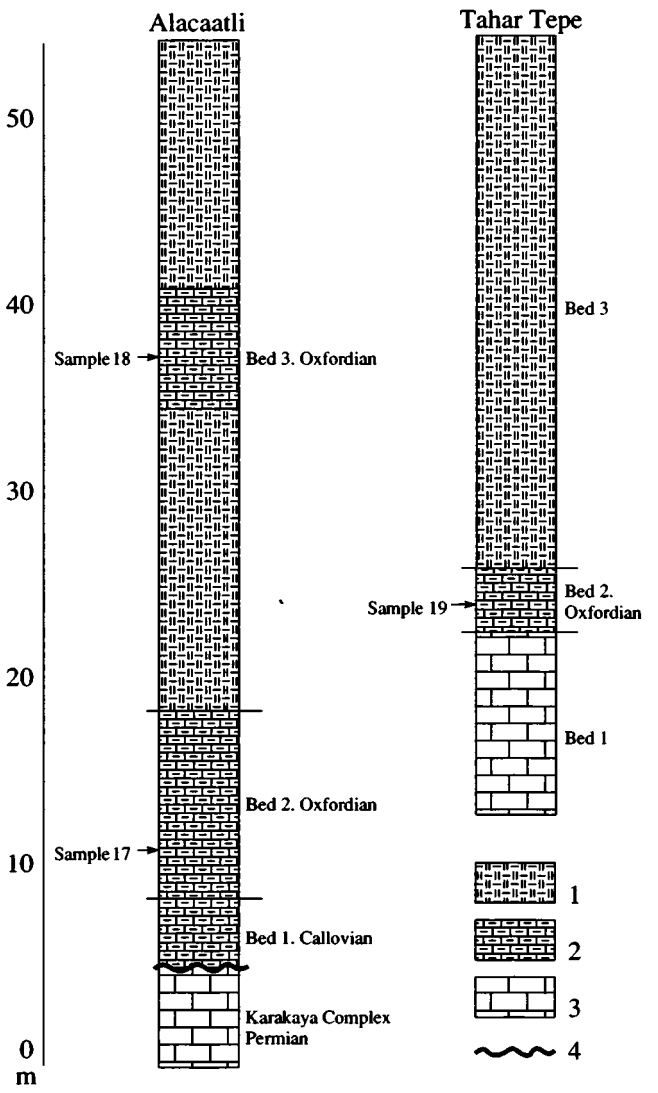


Fig. 3. Upper Jurassic sections of the Ankara area (bed numbers and their ages based on foraminifers are shown to the right of stratigraphic columns; sampling levels for foraminiferal analysis are shown to the left of the columns): (1) calcareous cherts; (2) cloddy limestones of the Ammonitico Rosso type; (3) massive and bedded limestones; (4) unconformity.

metamorphic schists (15%), and by relatively rare mica flakes. The cement is clayey-carbonate. The bed is composed of eroded products of the granite-metamorphic complex that accumulated in shallow-water coastal-marine settings under a high-energy influence of surf waves. The thickness is 20 m.

Bed 3: intercalated red cloddy sandy limestones, greenish gray flaggy sandstones, and red to gray marls. The red cloddy limestones contain 5–25% of silt- to sand-sized quartz clasts (0.07 to 0.35 mm across). Abundant organogenic detritus (10–30%) is represented by fragments of foraminiferal tests, molluscan shells, echinoderms, and algae. Ooliths and their fragments are rare. Flat detrital particles are oriented along bedding surfaces. The content of detrital material decreases upsection. The limestones can be classed with sandy-clayey-bioclastic wackestones and packstones that accumulated in relatively low-energy environments of moderately deep shelf near the interface of surf and tidal waves. Samples 2 and 3 (Sinemurian-Pliensbachian) represent these rocks. The lower part of the bed yields Sinemurian ammonoids *Asteroceras acceleratum* Hyatt, *A. blakei* (Spath), *Gleviceras* cf. *subquibalianum* (Pia), *G. victoris* (Dumortier), *Epideroceras (Coeloderoceras) roberti* (Hauer), *Paltehioceras elicitum* Buckman, and others (Bremer, 1965; Alkaya, 1998). The upper part of the bed encloses

ammonoids characterizing the Sinemurian-Pliensbachian transition and a lower Pliensbachian interval. These are *Phylloceras frondosum* (Reynes), *P. hebertinum* (Reynes), *Calliphylloceras* cf. *bicicolae* (Meneghini), *Epideroceras (Coeloderoceras) praecursor* (G. Geyer), *E. (C.) ponticum* (Pia), *Radstokiceras buvignieri* (d'Orbigny), *R. complanosum* (Simpson), *Phricodoceras taylori* (Sowerby), *Uptonia jamesoni* (Sowerby), and others (Bremer, 1965; Alkaya, 1998). The thickness is 10 m.

Bed 4: greenish gray to gray tuffaceous and silty mudstones with interbeds of coarse-grained polymictic sandstones. The clastic material in mudstones is represented by quartz and plagioclase grains and by small fragments of volcanics. The sandstones are composed of variably rounded unsorted quartz, quartzite, clayey limestone, and mudstone grains with subordinate fragments of metamorphic schists, acidic volcanics, and echinoderm remains. Rock clasts are slightly weathered and carbonatized. The bed differs from the underlying rocks owing to presence of acidic volcanics and sedimentary rocks among detrital material. These alternating sandstones and mudstones probably accumulated at the base of a low-angle slope or in the shelf-break zone under the influence of relatively high-energy descending currents. The thickness is 35 m.

Bed 5: greenish gray flaggy marls and clayey limestones with interbeds of yellowish gray sandstones. The limestones are composed of sparite and clastic admixture (up to 15%) represented by grains (0.1 to 0.2 mm across) of quartz, acidic volcanics, glauconite, mica flakes, and rare crystals of fresh hornblende. Organogenic detritus (up to 10%) includes thin-shelled bivalves, foraminifers, and echinoderm remains. Clasts and faunal remains are oriented along bedding surfaces. These sediments can be classed with sandy-clayey wackestones. The thickness is 15 m.

Bed 6: pink to red cloddy and flaggy sandy limestones consisting of micritic matrix engulfing grains (up to 10%) of quartz and altered glauconite. The notable admixture of ash is also characteristic of the bed. The content of bioclasts varies from 10 to 50%. They are represented by intact foraminiferal tests and their fragments, and by remains of thin-shelled bivalves, brachiopods, sponge spicules, echinoderms, and, probably, corals. Visible fucoids are replaced by calcite. The rocks can be classed with wackestones that formed in deep-water settings at some distance from a slope or in the shelf-break zone. The bed, which is characterized by Sample 7 (Toarcian), contains abundant ammonoids presumably of the late Domerian-early Toarcian age according to preliminary determination (Alkaya, 1998). The thickness is 2 m.

Bed 7: red to pink flaggy sandy marl deposited in sedimentation settings, which are similar, to a great extent, to those of underlying rocks. The thickness is 20 m.

Bed. 8: green flaggy siliceous-tuffaceous siltstones. The thickness is 3 m.

Bed 9: yellowish gray to dark gray flaggy siliceous-carbonate-tuffaceous siltstones and mudstones. Fine-grained clastic material includes grains of feldspar (mostly plagioclase), quartz, and volcanics (5% and less). The rocks are usually strongly altered, displaying presence of illite-chlorite paragenesis; their fresh varieties bear abundant vitroclasts. Some mudstones contain abundant siliceous or carbonate spherules (calcispherulids or radiolarians). This portion of the section is composed of alternating volcanomictic and tuffaceous siltstones and mudstones accumulated far from the seashore in a relatively deep-sea basin with volcanic activity. The thickness exceeds 100 m.

The Middle Jurassic section was studied and sampled in the Kosrelik Village area. Its description from the base upward is given below (Fig. 2).

Bed 1: gray or greenish to brownish gray thin-bedded carbonatized mudstones with thin interbeds of coarse-grained sandstones and organogenic-detrital limestones. The sandstones are composed of subangular to subrounded clasts of quartzites, micaceous-quartzose schists, gneisses, clayey sandstones, mudstones, algal and oolitic limestones, basaltic andesites, and echinoderm detritus. The thickness is 5 m.

Bed 2: red flaggy organogenic-detrital limestones containing angular echinoderm remains with algal crust, tests of different foraminiferal species, fragments of bivalves, brachiopods and ammonoids, and separate aptyichi. The bed is characterized by Sample 12 (Bajocian). The thickness is 2 m.

Bed 3: greenish gray siltstones with large siderite concretions. The thickness is 40 m.

Bed 4: red clayey micritic limestones and marls with remains of foraminifers, radiolarians, and planktonic (?) crinoids. The bed is exemplified by Sample 14 (Bajocian). Beds 2 and 4 of this section enclose the Bajocian ammonoid assemblage of *Eumedtoceras amplectens* Buckman, *E. amaltheiforme* (Vacek), *Fontannesia* cf. *whitehousei* Arkell, and *Sonninia crassispinata* Buckman (Alkaya, 1998). The thickness is 3 m.

Bed. 5: brownish, greenish, and yellowish gray mudstones with abundant small siderite concretions and rare vitroclasts. The apparent thickness is 30 m.

The Middle Jurassic sediments accumulated in deep-sea settings remote from the seashore and probably near the slope. The coarse-grained sandstones and floatstones from the lower part of the section are probably turbidites in origin. In general, the Lower-Middle Jurassic beds characterize a transgressive sedimentation trend, being deposited in very shallow-water nearshore settings in the Early Liassic and in relatively deep-sea environments in the Middle Jurassic. The interbeds of coarse-grained sediments within mudstones could be related to steep scarps and terraces at the slope. The clastic and organogenic-detrital material of different genesis was transported from the shallow-

water area by descending debris flows. In the Early Jurassic, eroded products of granites and metamorphic rocks were dominant and became gradually replaced by volcaniclastic material in the Middle Jurassic, when sedimentation occurred in the area with nearby volcanic activity.

The Upper Jurassic succession that has a tectonic contact with the Lower-Middle Jurassic one was studied in the Alacaatli and Tahar Tepe areas southwest of the Ankara center. The Upper Jurassic sediments rest here immediately upon the Permian-Triassic rocks of the Karakaya Complex. Below, we present description of the Alacaatli section (from the base upward) (Fig. 3).

Bed 1: light gray massive limestones of the Karakaya Complex (Permian).

Bed 2: red to brown-red clayey limestones and mudstones resting on the eroded Permian surface. Different blocks of the Permian limestones fill depressions and pockets in the erosional surface. The fine-grained sparite matrix of limestones encloses fragments of echinoderms, bivalves, foraminifers, and sponge spicules, and the rocks can be classed with wackestones. The bed is dated back to the Callovian based on ammonoids. The thickness is 2–5 m.

Bed 3: light yellowish gray cloddy limestones, which form extended lenses with a variable content of clay and organogenic-detrital material and show different recrystallization degree. The limestones exemplified by Sample 17 (Oxfordian) can mostly be classed with wackestones and packstones. The thickness is 10 m.

Bed 4: pinkish, yellowish, and greenish gray flaggy calcareous cherts with radiolarians and lenses of red to pinkish gray cloddy limestones; the rocks are characterized by Sample 18 (Oxfordian). The thickness is 35 m.

The Tahar Tepe section represents the succession of following beds (from the base upward; Fig. 3):

Bed. 1: light gray massive limestones. The thickness is 10 m.

Bed 2: gray and pinkish gray cloddy limestones with intact ammonoid and bivalve shells partly fragmented, ooliths, lithoclasts of micritic limestones, and rounded coral remains. These rocks exemplified by Sample 19 (Oxfordian) are classed with packstones and grainstones. The thickness is 2–4 m.

Bed 3: pink, gray, and yellow calcareous cherts with radiolarians. The apparent thickness is 25–30 m.

As a whole, the Callovian-Upper Jurassic sediments of the Ankara area accumulated in the open sea basin. They exemplify a transgressive trend of sedimentation first in shelf environments during the Callovian and then in deeper hemipelagic settings during the Late Jurassic.

INVESTIGATION METHODS

In the Ankara area, foraminifers occur mainly in carbonate rocks, and samples were collected from all limestone members and separate beds in studied sections. Some additional samples taken from other lithological types, mostly from fine-grained terrigenous rocks, are lacking foraminifers. To the contrary, carbonate varieties yielded abundant microfossils from many stratigraphic levels. In total, 19 samples were collected from the Sinemurian-Oxfordian interval of the section.

It should be noted that the study of foraminifers from the Jurassic sections of central Turkey was difficult because of the rock composition and properties. Compact silicified and clayey limestones of the Ammonitico Rosso facies cannot be disintegrated by traditional methods. Inasmuch as the study of foraminifers in thin sections is inappropriate for complete characterization of their composition and morphology, every sample was divided in two portions. One part was used to prepare thin sections (8–10 thin sections per a sample), and another one was treated in the formic acid with subsequent washing out of unsolved residue and picking up of foraminiferal tests. Noteworthy is that the compositions of foraminiferal assemblages determined in thin sections and residues appeared to be different: benthic foraminifers of the Ammodiscidae, Lituolidae, and Nodosariidae families prevailed in thin sections while residues contained representatives of the Spirillinidae, Ataxophragmiidae, Epistominidae, and Involutinidae and also planktonic forms. Species of the Nodosariide family, which have calcite tests, were represented only by rare casts because of the complete dissolution of their tests.

The SEM photographs of extracted tests were made by N.V. Gor'kova in the Laboratory of Physical Methods at the Geological Institute RAS. Thin sections with foraminifers were photographed under microscope by Bragin and Kuznetsova from the Laboratory of Micropaleontology of the same institute, where the laboratory treatment of all samples was carried out as well.

COMPOSITION, DIVERSITY, AND STRATIGRAPHIC DISTRIBUTION OF JURASSIC FORAMINIFERS IN THE ANKARA AREA

General Characteristics

Foraminifers occur in all studied sections (Figs. 1–3), within the stratigraphic interval from the Sinemurian to Oxfordian, i.e., through all three series of the Jurassic System. Representative foraminiferal assemblages are found in 13 samples of Sinemurian, Pliensbachian, Toarcian, Bajocian, and Oxfordian rocks. Foraminifers have not been encountered in the Bathonian, Aalenian, and Callovian sediments. The foraminiferal assemblages include representatives of 13 orders, 57 genera, and approximately 120 species (Table 1, Plates I–V). We used the foraminiferal classification accepted in Russia (Maslakova, 1990; Maslakova and Gorbachik, 1995), according to which the class Foraminifera consists of 8 subclasses and 29 orders. Some genera such

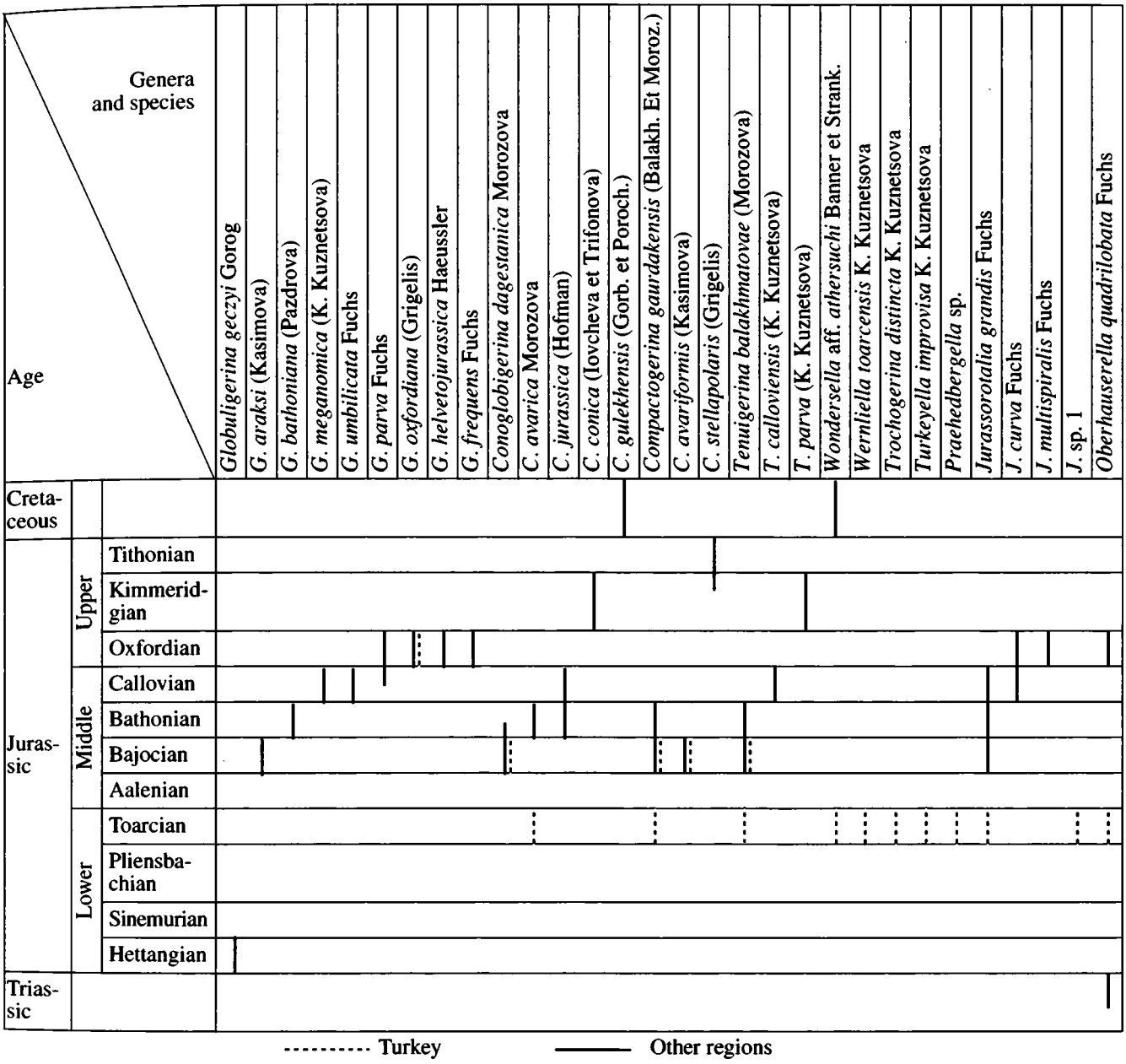
Table 1. Stratigraphic distribution of foraminifers in Jurassic sediments of central Turkey

Sample number	2, 3, 4	7		12, 14			17, 18, 19
Stage / Genera and species	Sinemurian-Pliensbachian	Toarcian	Aalenian	Bajocian	Bathonian	Callovian	Oxfordian
Ammodiscus asper (Terquem)							
A. septentrionalis Gerke							
A. siliceus (Terquem)							
A. pseudoinfimus Gerke et Sossipatrova							
Ammobaculites sp.							
Dentalina torta Terquem							
D. pseusocommunis Franke							
D. exilis Franke							
D. obscura Terquem							
D. sinemuriensis Terquem							
D. ex. gr. *communis* d'Orbigny							
Nodosaria turgida Gerke							
N. variabilis Terquem et Berthelin							
N. mitis Terquem et Berthelin							
N. nitidana Brand							
N. claviformis Terquem							
N. columnaris Franke							
Pseudonodosaria dea Schleifer							
P. ventrosa Schleifer							
P. vulgata Bornemann							
Lingulina tenera tenera Bornemann							
Geinitzinita sp.							
Ichtyolaria sp.							
Frondicularia sp.							
Frondiculinita dubiella (Gerke)							
Lenticulina acutiangulata Terquem							
Astacolus praefoliaceus (Gerke).							
Ophthalmidium leischneri (Kristan-Tollmann)							
O. martanum (Farinacci)							
Praeophthalmidium postconcentricus (Kasim.)							
Trocholina turris Frentzen							
Rhizammina sp.							
Saccammina inanis Gerke et Soss.							
Hyperammina labaensis Ant.							
Hippocrepina liassica (Brand)							
Lituotuba sp.							
Scherochorella sp.							
Ammobaculites vetustus (Terq. et Berth.)							
A. fontinensis (Terquem)							
Reophax torus Crespin							
R. scorpiurus Mont.							
Textulariopsis sp.							
Gaudryina topagorukensis Tappan							
Trochammina squamatiformis Kapt.							
Verneuilinoides tchegemensis Mak.							
Martinotiella sp.							
Coscinophragma sp.							
Lagena ovata Terquem							
Geinitzinita octocostata (Bartenstein et Brand)							
Lenticulina vetusta (d'Orbigny)							
Planularia sp.							

Table 1. (Contd.)

Sample number	2, 3, 4	7		12, 14			17, 18, 19
Stage / Genera and species	Sinemurian-Pliensbachian	Toarcian	Aalenian	Bajocian	Bathonian	Callovian	Oxfordian
Citharina clathrata (Terquem)							
Cornuspira orbicula (Terquem et Berthelin)							
Repmanina charoides (Jones et Park.)							
Praeophthalmidium mamontovae (Ant.)							
Ophthalmidium infraoolithicum (Terquem)							
Bullopora rostrata Quenst.							
Wernliella toarcensis K. Kuzn.							
Turkeyella improvisa K. Kuzn.							
Trochogerina distincta K. Kuzn.							
Wondersella aff. *athersuchi* Bann. et Strank							
Jurassorotalia grandis Fuchs							
J. multispiralis Fuchs							
Oberhauserella quadrilobata Fuchs							
Ammodiscus infimus Issler							
Budaschevaella sp.							
Haplophragmoides sp.							
Evolutinella convexa (Ant.)							
Ammobaculites agglutinans (d'Orbigny)							
Dentalina matutina d'Orbigny							
Nodosaria metensis Terquem							
N. variabilis Terquem et Berthelin							
Lenticulina mironovi Dain							
Globuligerina araksi (Kasim).							
Conoglobigerina dagestanica (Moroz.)							
C. avarica (Moroz.)							
Compactogerina avariformis (Kasim.)							
C. qaurdakensis (Balakh. et Moroz.)							
Tenuigerina balakhmatovae (Moroz.)							
Conorboides paraspis (Schwag.)							
Placopsilina argoviensis Haeusler							
Paleogaudryina varsoviensis (Bielecka et Pozarysky)							
Cornuspira nitervacare Azbel							
Spirillina helvetica Kubler et Zwingli							
S. polygyrata Gumbel							
Trocholina transversarii Paalzow							
Eoguttulina simplex (Terquem)							
Ophthalmidium strumosum (Gumbel)							
O. marginatum (Wisniowski)							
O. stuifense (Paalzow)							
O. sp.							
Ophthalmidium sagittum (E. Byk.)							
Labalina milioliniforme (Paalzow)							
Miliolinella frumenta (Azbel)							
Nubeculinella epistominis Dain							
N. megaspiralis Danitch							
Lenticulina belorussica (Mitjanina)							
L. jakovlevae Azbel							
Epistomina nemunensis Grigelis							
E. multialveolata Grigelis							
E. gracilis Dain							
Epistomimita formastra Azbel							
?Kaptarenkoella sp.							
Paulina furssenkoi Grigelis							
Globuligerina oxfordiana (Grigelis)							

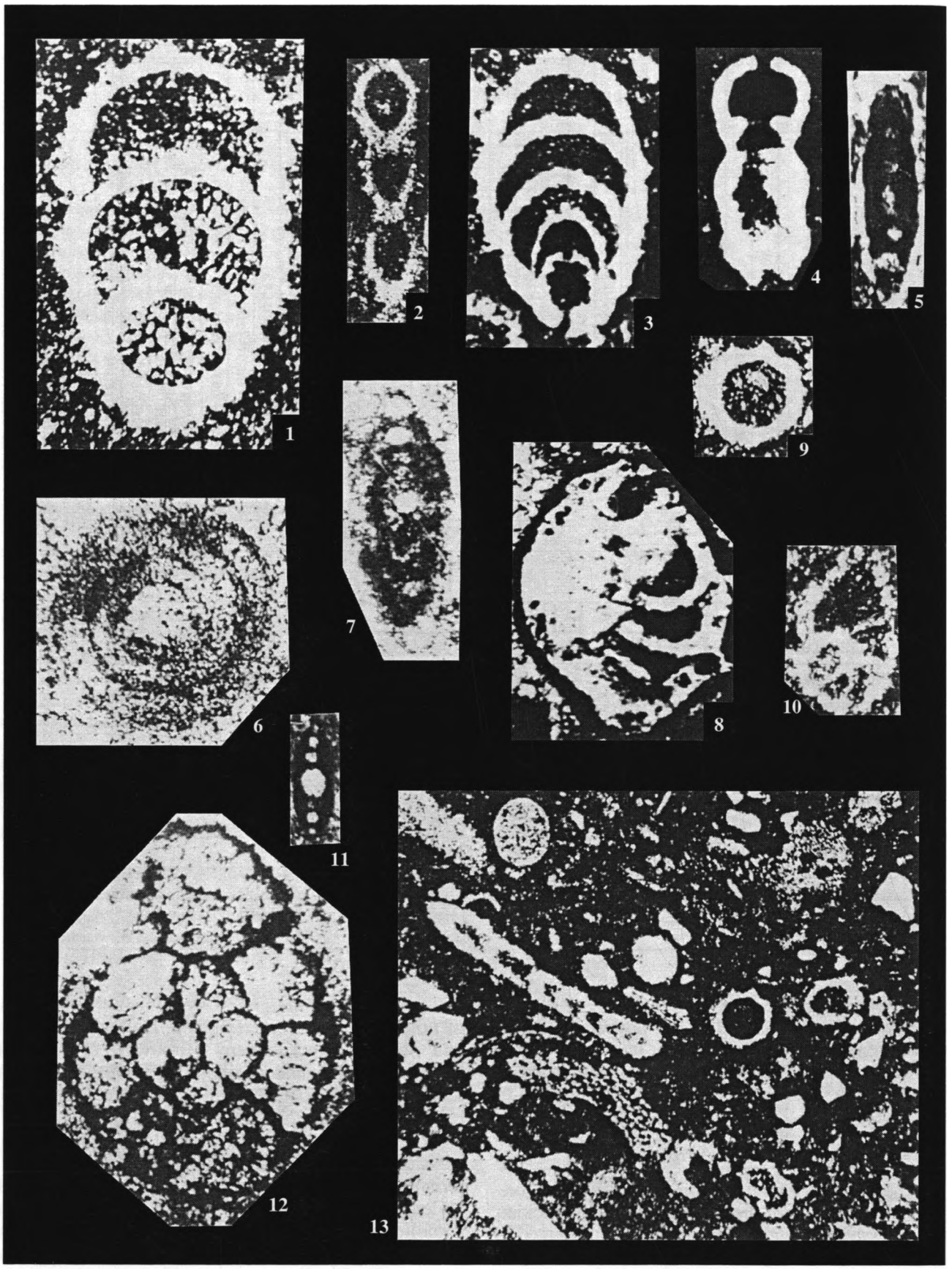
Table 2. Stratigraphic distribution of planktonic foraminifers in Jurassic sections



as *Dentalina, Lingulina*, and *Pseudonodosaria* (family Nodosariidae) are considered sensu lato without their subdivision into recently proposed genera, such as *Mesodentalina, Prodentalina*, and others.

The Lower Jurassic

In the Lower Jurassic portion of the sections, foraminifers are present in the Sinemurian, Pliensbachian, and Toarcian sediments, where they are represented by solely cosmopolitan species or by taxa of wide geographic distribution. The latter are known from the Lias of northwestern Europe: England, Holland, Denmark, and northwestern Germany (Brouwer, 1969), from Swiss and Lithuania (Grigelis and Norling, 1999), northern Siberia (Gerke, 1961), western Carpathians (Gazdzicki, 1983), North Caucasus (Makar'eva, 1971, 1982); Portugal (Exton, 1979), and Morocco (Sosipatrova and Rohhali, 1974). This is an important characteristic of studied assemblages lacking the Tethyan endemic taxa, the so-called "larger foraminifers" of the Lituolidae, Cyclamminidae, and Pfenderinidae families, which have intricate tests with the alveolarian-labyrinth walls. The Tethyan endemic species characteris-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All foraminiferal specimens in Plates I–V are from Jurassic sediments of central Turkey. Localization of sections and samples is shown in Figs. 1–3. Collections nos. 4772 and 4775 are stored at the Geological Institute RAS, Moscow

Plate I. Lower Jurassic benthic foraminifers (except for Figs. 8 and 12).

(1) *Nodosaria nitidana* Brand, Specimen 4775/1, Kosrelik Kizigi section, Sample 3, Sinemurian-Pliensbachian, longitudinal section, ×130. (2) *Nodosaria claviformis* Terquem, Specimen 4775/2, Kosrelik Kizigi section, Sample 2, Sinemurian-Pliensbachian, longitudinal section, ×70. (3) *Geinitzinita* sp., Specimen 4775/3, Kosrelik Kizigi section, Sample 2, Sinemurian-Pliensbachian, longitudinal section, ×160. (4) *Geinitzinita* sp., Specimen 4775/4, Kosrelik Kizigi section, Sample 7, Toarcian, longitudinal section, ×180. (5) *Ophtalmidium leischneri* (Kristan-Tollmann), Specimen 4775/5, Kosrelik Kizigi section, Sample 3, Sinemurian-Pliensbachian, longitudinal section, ×160. (6, 7) *Ophtalmidium* cf. *martanum* (Farinacci): (6) Specimen 4775/6, Kosrelik Kizigi section, Sample 3, Sinemurian-Pliensbachian, equatorial section, ×300; (7) Specimen 4775/7, the same locality and age, transverse section, ×180. (8) *Lenticulina* cf. *belorussica* (Mitjanina), Specimen 4775/8, Tahar Tepe section, Sample 19, Oxfordian, tangential section, ×50. (9) *Nodosaria columnaris* Franke, Specimen 4775/9, Kosrelik Kizigi section, Sample 2, Sinemurian-Pliensbachian, transverse section, ×360. (10) *Astacolus praefoliaceous* (Gerke), Specimen 4775/10, Kosrelik Kizigi section, Sample 2, Sinemurian-Pliensbachian, longitudinal section, ×140. (11) *Praeophtalmidium* cf. *postconcentricus* (Kasimova), Specimen 4775/11, Kosrelik Kizigi section, Sample 7, Toarcian, transverse section, ×150. (12) *Ammobaculites* sp., Specimen 4775/12, Kosrelik Kizigi section, Sample 12, Bajocian, equatorial section, ×350. (13) sandy-clayey bioclastic micritic wackestone with remains of bivalves, echinoderms, and foraminifers, Kosrelik Kizigi section, Sample 2, Sinemurian-Pliensbachian, ×200.

tic of shallow-water assemblages are known from the Lower-Middle Jurassic sections of the Antalya area of Turkey (Bassoullet and Poisson, 1975). Another remarkable peculiarity of Early Jurassic foraminiferal assemblages is association of benthic forms with diverse planktonic species first discovered in the Toarcian sediments of the Kosrelik Kizigi section. The planktonic foraminifers belong to 10 genera and 11 species, three of which were first described from this locality (Kuznetsova, 2002). Other genera and species have been detected in the Toarcian strata. Previously, they were reported from younger Jurassic and Cretaceous sediments of the Crimea, Caucasus, Central Asia, West Europe, and United Arab Emirates (Table 2) (Morozova and Moskalenko, 1961; Banner and Strank, 1987; Boudagher-Fadel *et al.*, 1997; Fuchs, 1973, 1975). The Liassic assemblages are characterized by a distinct prevalence of Nodosariidae genera *Nodosaria, Dentalina, Lagena, Lingulina, Pseudonodosaria, Ichtyolaria, Frondicularia, Lenticulina, Astacolus, Geinitzinita*, and *Citharina*. These are accompanied by agglutinated (genera *Ammodiscus, Trochammina, Glomospira*, and *Ammobaculites*) and calcareous forms (genera *Trocholina, Praeophtalmidium* (=*Cornuloculina*), and *Ophtalmidium*) representing other families. The last two genera were observed in abundance in thin sections, and some forms were identified at the species level (Tables 1, 2).

Sinemurian and Pliensbachian Stages

Foraminifers of these stages were studied in samples from red cloddy sandy limestones of the Ammonitico Rosso facies in the Kosrelik Kizigi section (Fig. 2). The assemblage from Samples 2–4 includes the following characteristic species: *Ammodiscus* cf. *asper* (Terquem), *A. septentrionalis* Gerke, *A. siliceous* (Terquem), *Lingulina tenera tenera* Bornemann, *Nodosaria nitidana* Brand, *N. mitis* (Terquem et Berthelin), *N.* cf. *claviformis* Terquem, *N. columnaris* Franke, *Dentalina sinemuriensis* Terquem, *D. obscura* Terquem, *D. exilis* Franke, *Ichtyolaria* sp., *Trocholina turris* Frentzen, *Ophtalmidium* cf. *martanum* (Farinacci), and *O. leischneri* (Kristan-Tollmann) (Table 1). The assemblage composition allows the host sediments to be dated back to the Sinemurian-Pliensbachian. The more accurate dating is impossible so far, because foraminifers are studied only in thin sections (26 in total), and no tests were extracted from the rocks after the acidic treatment. The taxa identified in thin sections had a long existence period being common in both the Sinemurian and Pliensbachian stages. This conclusion is

Plate II. Benthic foraminifers from the Toarcian Stage of central Turkey (Kosrelik Kizigi section, Sample 7).

(1) *Scherochorella* sp., Specimen 4775/13, side view, ×115. (2) *Saccammina* sp., Specimen 4775/14, side view, ×115. (3) *Reophax scorpirus* Montfort, Specimen 4775/15, remains of the internal layer after acidic treatment of the test, side view, ×115. (4a, 4b, 5) *Reophax torus* Crespin: (4) Specimen 4775/16, (4a) test without last chamber, ×115; (4b) a fragment of test surface, ×2000; (5) Specimen 4775/17, side view. (6, 7) *Ammodiscus siliceous* (Terquem), Specimens 4775/68 and 4775/18, side view, ×115. (8, 9) *Ammodiscus asper* (Terquem), Specimens 4775/19 and 4775/20: (8) side view, (9) peripheral view, ×115. (10) *Ammodiscus fontinensis* (Terquem), Specimen 4775/21, side view, ×115. (11, 16) *Ammodiscus vetustus* (Terquem et Berthelin): (11) Specimen 4775/22, side view, ×115; (16) Specimen 4775/23, side view, ×115. (12) *Gaudryina topagorukensis* Tappan, Specimen 4775/24, side view, ×70. (13) *Martinottiella* sp., Specimen 4775/25, side view with the initial tetraserial part passing immediately into uniserial part, ×70. (14) *Textulariopsis* sp., Specimen 4775/26, side view, ×70; (15) *Verneuilinoides tchegemensis* Makarjeva, Specimen 4775/27, side view with the partly visible apertural surface, ×115. (17) *Saccammia inanis* Gerke et Sossipatrova, Specimen 4775/28, apertural surface, ×115. (18–20) *Trochammina squamatiformis* Kaptarenko, Specimens 4775/29–4775/31: (18, 19) opposite sides; (20) peripheral side, ×115. (21) *Repmanina charoides* (Jones et Parker), Specimen 4775/32, side view with an open end of the tubular chamber, ×115. (22) *Hyppocrepina liassica* (Brand), Specimen 4775/33, side view, ×115. (23) *Hyperammina labaensis* Antonova, Specimen 4775/34, side view, ×115. (24) *Rhizammina* (?) sp., Specimen 4775/35, ×35.

Plate III. Benthic foraminifers from the Bajocian Stage of central Turkey (Kosrelik section).

(1) *Reophax torus* Crespin, Specimen 4775/36, Sample 12, side view, ×115. (2–5) *Verneuilinoides tchegemensis* Makarjeva, Sample 12: (2) side view and (3) apertural surface of Specimen 4775/37; (4) Specimen 4775/38, apertural surface; (5) Specimen 4775/39, side view, ×115. (6) *Evolutinella convexa* (Antonova), Specimen 4775/40, Sample 12, side view, ×115. (7, 8) *Budaschevaella* sp. Specimen 4775/27, Sample 14: (7) side and (8) peripheral views, ×115. (9) *Ammobaculites agglutinans* (d'Orbigny), Specimen 4775/42, Sample 14, side view, ×115. (10–12) *Repmanina charoides* (Jones et Parker), Specimens 4775/43–4775/45, Sample 12: (10) ×240; (11) side view with an open end of the tubular chamber, ×115; (12) ×115. (13, 14, 17, 18) *Conorboides paraspis* (Schwager): (13) Specimen 4775/46, spiral side; (14) Specimen 4775/47, umbilical side, ×115; (17, 18) Specimen 4775/48, Sample 14: (17) spiral side, ×240, (18) a fragment of test surface, ×2000. (15, 16) *Ammobaculites* aff. *fontinensis rossica* Makarjeva, Specimens 4775/49 and 4775/50, Sample 12: (15) side view; (16) peripheral view, ×115.

well consistent with the rock age inferred from the ammonoid assemblage (Alkaya, 1998).

Toarcian Stage

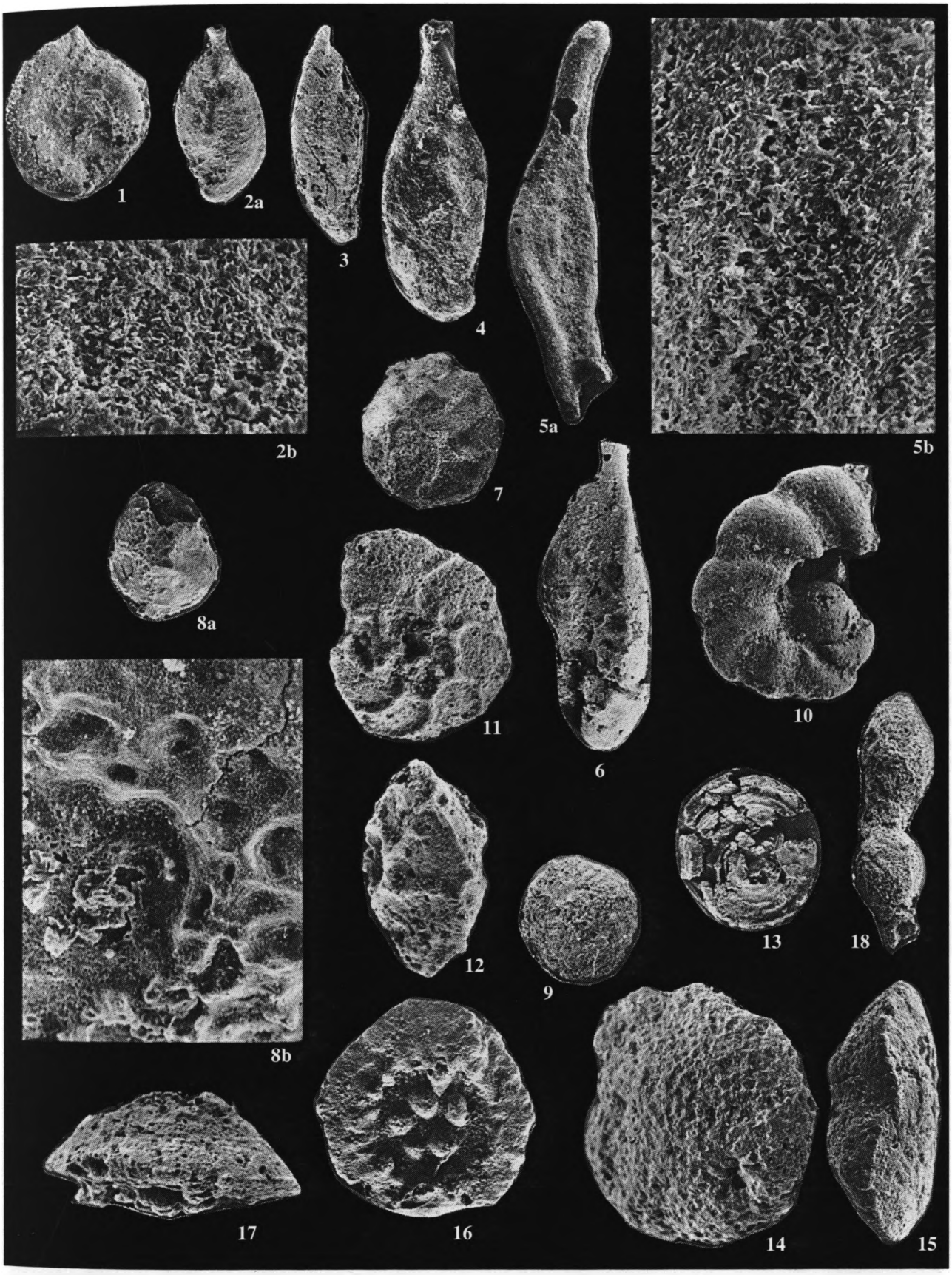
The reddish sandy limestones of the Ammonitico Rosso type and marls (Sample 7) from a higher portion of the section enclose an extremely diverse assemblage of abundant foraminifers. Thin sections demonstrate prevalence of Nodosariidae genera *Lenticulina, Citharina, Nodosaria*, and *Dentalina* accompanied by rare *Ammodiscus, Trocholina, Cornuspira*, and *Praeophtalmidium* forms (Table 1). Species of the last genus are represented by typical Toarcian species *Praeophtalmidium* cf. *postconcentricum* (Kasimova) and *P. mamontovae* (Antonova) described from the Toarcian sediments of the North Caucasus (Antonova, 1975; Temirbekova and Antonova, 1985). In addition, Sample 7 contains Toarcian *Citharina clathrata* (Terquem) and *Ophtalmidium infraoolithicum* (Terquem), and these taxa prove the age estimate based on ammonoids.

The foraminiferal tests extracted from the rocks are more diverse and represent characteristic species first appearing in the Lower Jurassic section. These are *Verneuilinoides tchegemensis* Makarjeva described from the Toarcian sediments of the northeastern Caucasus (Makar'eva, 1971, 1982), *Trochammina squamatiformis* Kaptarenko peculiar for the Toarcian sequences of the Dnieper-Donets depression, *Hyppocrepina liassica* (Brand) widespread in Toarcian foraminiferal assemblages of the North Caucasus, *Textulariopsis* sp. registered in the Toarcian rocks of Gibraltar (Boudagher-Fadel *et al.*, 2001), *Ammobaculites vetustus* (Terquem et Berthelin) described from the Liassic deposits of the Parisian Basin and occurring in the Toarcian sediments of the North Caucasus, and others. The same interval is also marked by the first appearance of *Repmanina charoides* (Jones et Park.) known from the Toarcian of the North Caucasus and *Hyperammina labaensis* Antonova typical of the Toarcian and Aalenian sediments in the same region (Antonova, 1958). In total, Toarcian foraminifers from Turkey are represented by 40 species, thus being more diverse than the Bajocian and Oxfordian assemblages.

The most remarkable feature of the Toarcian foraminiferal community is, however, the appearance of planktonic species, as it is established in the Kosrelik Kizigi section (Sample 7). It should be noted that it was Wernli (1988, 1995) who first mentioned finds of Toarcian planktonic foraminifers in Turkey. He illustrated sections of forms referred to as "protoglobigerinids" and described also representatives of the *Oberhauserella* genus from the Toarcian sediments of Switzerland. The assemblage of planktonic foraminifers from the Kosrelik Kizigi section includes 10 genera and 11 species. Three species *Wernliella toarcensis* K. Kuzn., *Turkeyella improvisa* K. Kuzn., and *Trochogerina distincta* K. Kuzn. were first described from this locality (Kuznetsova, 2002). They associate with *Wondersella* aff. *athersuchi* Banner et Strank described from the Lower Cretaceous sediments of the United Arab Emirates (Banner et Strank), *Jurassorotalia grandis* Fuchs, *Oberhauserella quadrilobata* Fuchs (Fuchs, 1973, 1975), and species of genera *Conoglobigerina, Compactogerina*, and *Tenuigerina* known from the Middle Jurassic of the North Caucasus, Central Asia, and

Plate IV. Benthic foraminifers from the Oxfordian Stage of central Turkey.

(1) *Ophtalmidium* sp., Specimen 4775/51, Tahar Tepe section, Sample 19, side view, ×115; (2, 3, 6) *Ophtalmidium strumosum* (Gumbel), Tahar Tepe section, Sample 19: (2) Specimen 4775/52, (2a) side view, ×115; (2b) fragment of the test surface, ×2000; (3) side view, ×115; (6) side view, ×115. (4) *Ophtalmidium marginatum* (Wisniowski), Specimen 4775/55, Alacaatli section, Sample 18, side view, ×115. (5) *Ophtalmidium sagittum* (E. Bykova), Specimen 4775/56, Tahar Tepe section, Sample 19: (5a) side view, ×115; (5b) a fragment of test surface, ×2000. (7) *Epistomina gracilis* Dain, Specimen 4775/57, Alacaatli section, Sample 18, side view, ×115. (8) *Epistominita formastra* Azbel. Specimen 4775/58, Tahar Tepe section, Sample 19: (8a) umbilical side, ×115, (8b) fragment of the test surface, ×2000. (9, 14) *Epistomina multialveolata* Grigelis, Tahar Tepe section, Sample 19: (9) Specimen 4775/59, umbilical side, ×115; (14) Specimen 4775/60, spiral side, ×240. (10) *Lenticulina* cf. *jakovlevae* Azbel. Specimen 4775/61, Tahar Tepe section, Sample 19, cast with a dissolved calcite wall, side view, ×115. (11, 12) *Epistomina nemunensis* Grigelis, Specimen 4775/62, Tahar Tepe section, Sample 19, ×115: (11) spiral side, (12) peripheral view. (13) *Spirillina helvetica* Kublter et Zwingli, Specimen 4775/63, Alacaatli section, Sample 18, side view, ×115. (14, 15) *Kaptarenkoella* (?) sp. Specimen 4775/64, Tahar Tepe section, Sample 19, ×270: (14) umbilical side, (15) peripheral view. (16, 17) *Trocholina transversarii* Paalzow, Specimen 4775/65, Tahar Tepe section, Sample 19, ×115: (16) umbilical side, (17) peripheral view. (18) *Nubeculinella megaspiralis* Danitsch, Specimen 4775/66, Tahar Tepe section, Sample 19, a fragment of the free test part, ×115.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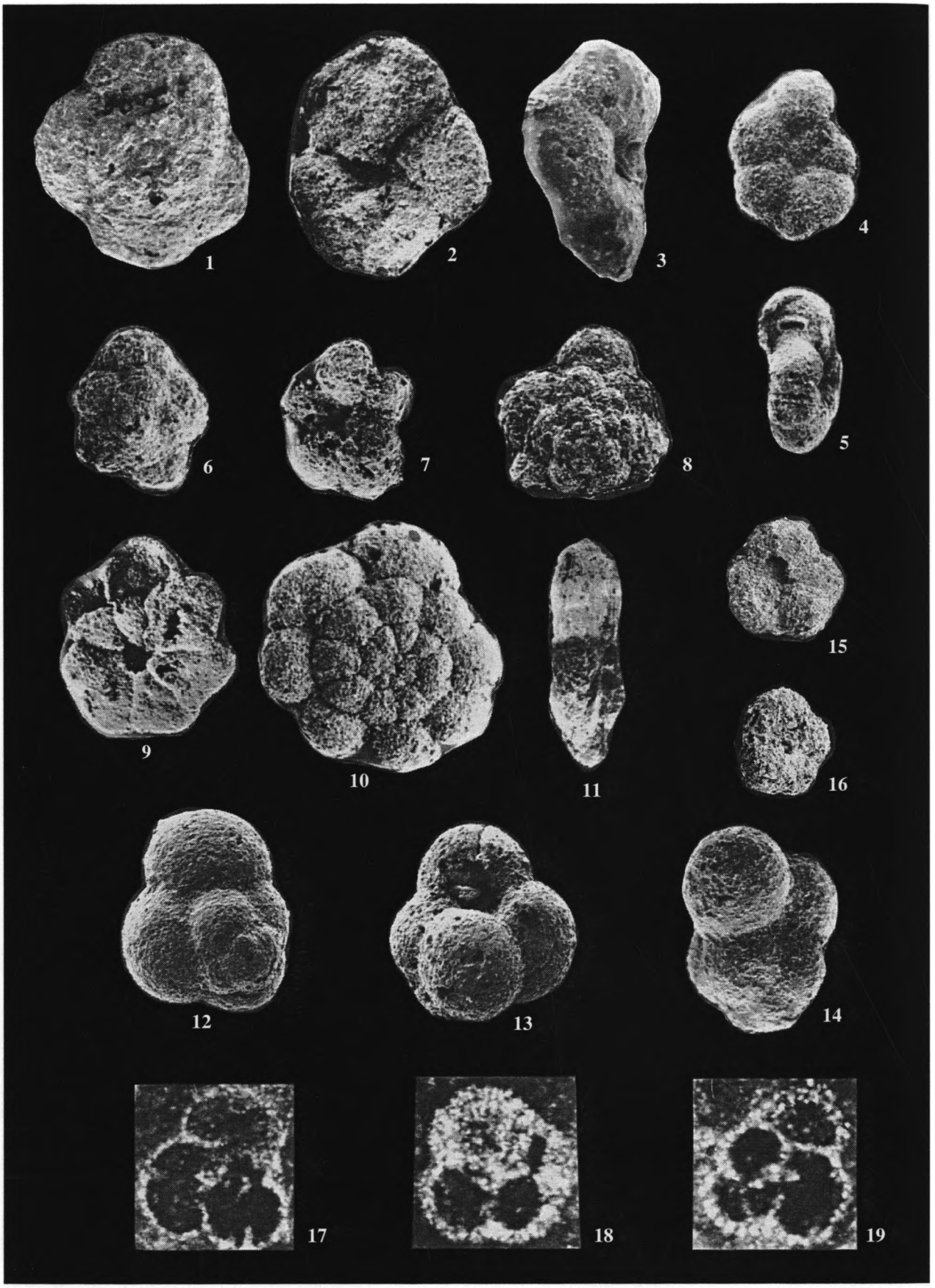
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southeastern Europe (Plates I, V). These species are mentioned and pictured in the English monograph devoted to planktonic foraminifers (Boudagher-Fadel *et al.*, 1997).

Middle Jurassic, Bajocian Stage

The Middle Jurassic foraminifers were found in interbeds of Ammonitico Rosso limestones in the lower and middle parts of the Kosrelik section (Samples 12, 14) (Fig. 2). Their assemblage substantially differs from the Liassic one, although some species such as *Verneuilinoides tchegemensis* Mak., *Repmanina charoides* (Jones et Park.), and *Reophax torus* Crespin are inherited from the Toarcian sediments. It is remarkable that planktonic species of the *Globuligerina, Conoglobigerina, Tenuigerina*, and *Compactogerina* genera are components of the Middle Jurassic foraminiferal assemblage. These genera were described from the Bajocian sediments of the northeastern Caucasus and Turkmenistan (Morozova and Moskalenko, 1961; Gorbachik and Kuznetsova, 1997) and considered previously as characteristic only of these two regions. Recently, *"Protoglobigerina"* forms were discovered in the lower Bajocian sediments of Morocco and Hungary (Wernli et Gorog, 1999). In Turkey, Bajocian planktonic foraminifers were unknown before. Presence of planktonic taxa was noted here only in the Callovian (Altiner, 1991) and younger sediments. In the assemblage under consideration, planktonic forms are associated with abundant benthic foraminifers of the *Reophax, Ammodiscus, Ammobaculites, Budaschevaella, Evolutinella, Verneuilinoides, Lenticulina, Citharina, Nodosaria*, and *Frondicularia* genera, which are of wide stratigraphic and geographic ranges. They are registered in the Bajocian sequences of northwestern Europe, Morocco, Near East, the Caucasus, and West Siberia (Grigelis and Norling, 1999; Gerke, 1961; Sossipatrova and Rohhali, 1974; Kuznetsova *et al.*, 1996). It is also important to note the appearance of ceratobuliminids represented by scarce specimens of *Conorboides paraspis* (Seibold). Based on presence of planktonic species *Conoglobigerina dagestanica* (Morozova), *Compactogerina gaurdakensis* (Balakhmatova et Morozova), and *Tenuigerina balakhmatovae* (Morozova) identified in thin sections and macerates, the age of host rock is determined as the Bajocian.

Upper Jurassic, Oxfordian Stage

Sediments of the Oxfordian stage are the youngest in examined sections of Central Turkey, where we studied foraminifers. The Oxfordian foraminiferal assemblages are found in the Alacaatli and Tahar Tepe sections (Samples 17–19) where they are confined to interbeds and lenses of pink and yellowish gray limestones of the Ammonitico Rosso type (Fig. 3). These rocks yield diverse and abundant benthic and planktonic foraminifers representing cosmopolitan taxa. The planktonic group includes *Globuligerina oxfordiana* (Grigelis), which forms mass accumulations in some samples (approximately 1000 specimens and more). A. Azbel' (see Mesezhnikov *et al.*, 1989) who studied Oxfordian foraminifers of the East European platform also noted an extremely high abundance of this taxon. *Globuligerina oxfordiana* (Grigelis) is a sole planktonic taxon of Oxfordian foraminifers that is of a distinct and narrow stratigraphic range (lower to middle Oxfordian) and of extremely wide geographic distribution from Sweden and Baltic areas to the Crimea, Caucasus, Central Asia, Italy, Turkey, Syria, and Ethiopia. The remarkable coexistence of benthic and planktonic forms in the Oxfordian foraminiferal assemblage was previously noted in the Alacaatli section of central Turkey (Gorbachik and Kuznetsova, 1997; Kuznetsova and Bragin, 1999). New data presented here supplement the former characterization of this community that includes zonal index species characteristic of the lower-middle Oxfordian sediments. The wide spatial distribution of species imparts a substantial value to this assemblage for subglobal correlation.

Other characteristic Oxfordian forms are *Globuligerina oxfordiana* (Grigelis), *Ophtalmidium sagittum* (E. Byk.), *O. strumosum* (Gumbel), *Epistomina nemunensis* Grig., *E. multialveolata* Grig., *Epistominita formastra* Azbel, *Paulina furssenkoi* Grig., and *Trocholina transversarii* Paalzow (Plate IV). The data on the early-middle Oxfordian age of their host beds are consistent

Plate V. Planktonic foraminifers from Jurassic sections of central Turkey.

(1–3) *Wernliella toarcensis* K. Kuznetsova, Holotype 4772/6, central Turkey, Kosrelik Kizigi section, Sample 7, Toarcian Stage, ×115: (1) spiral side, (2) umbilical side, (3) peripheral view. (4, 5) *Turkeyella improvisa* K. Kuznetsova, Holotype 4772/5, central Turkey, Kosrelik Kizigi section, Sample 7, Toarcian Stage, ×115: (4) side view (biumbilical test), (5) peripheral view with visible septal subbasal aperture surrounded by a lip. (6–8) *Trochogerina distincta* K. Kuznetsova, paratype 4775/67, central Turkey, Kosrelik Kizigi section, Sample 7, Toarcian Stage, ×115: (6) spiral side, (7) umbilical side, (8) holotype 4772/8, spiral side of the microspherical test. (9–11) *Jurassorotalia grandis* Fuchs, Central Turkey, Kosrelik Kizigi section, Sample 7, Toarcian Stage, ×115: (9) Specimen 4772/17, umbilical side; (10) spiral side and (11) peripheral view of Specimen 4772. (12–14) *Globuligerina oxfordiana* (Grigelis), Specimens 4775/70–4775/72, central Turkey, Alacaatli section, Sample 19, Oxfordian Stage, ×180: (12) spiral side, (13) umbilical side, (14) side view. (15) *Compactogerina gaurdakensis* (Balakhmatova et Morozova), Specimen 4772/27, central Turkey, Kosrelik section, Sample 14, Bajocian Stage, umbilical side, ×115. (16) *Tenuigerina balakhmatovae* (Morozova), Specimen 4775/74, central Turkey, Kosrelik section, Sample 7, Toarcian Stage, umbilical side, ×115. (17) *Tenuigerina balakhmatovae* (Morozova), Specimen 4772/22, central Turkey, Kosrelik section, Sample 14, Bajocian Stage, equatorial section, ×270. (18) *Compactogerina gaurdakensis* (Balakhmatova et Morozova), Specimen 4772/25, central Turkey, Kosrelik section, Sample 14, Bajocian Stage, equatorial section, ×270. (19) *Conoglobigerina dagestanica* (Morozova), Specimen 4772/26, central Turkey, Kosrelik section, Sample 14, Bajocian Stage, equatorial section, ×270.

with age estimates based on radiolarians (Bragin and Tekin, 1999).

CONCLUSIONS

Foraminifers in the Jurassic sections exposed in the Ankara area are abundant and diverse mainly in facies of relatively deep-sea settings. Their assemblages are characterized by prevalence of cosmopolitan taxa, which are of extremely wide geographic ranges. Beginning from the Toarcian Stage and upsection, diverse planktonic forms persistently appear among benthic foraminifers. The data obtained indicate that foraminiferal assemblages should be used for detailed subdivision and correlation of Jurassic sections in central Turkey. Moreover, it is of interest to continue investigation of cosmopolitan foraminiferal assemblages in Tethyan sections, because they bear a high potential for solving the problem of the Boreal-Tethyan correlation of the Jurassic deposits.

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Palynological Characteristics of Paleocene Deposits in the Northern West Siberian Plate

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Abstract—Information on Paleocene deposits of Western Siberia is scarce and fragmental. The palynological analysis of new materials is important therefore, and the data obtained can be used in stratigraphy and for interregional correlation and reconstruction of paleoenvironments of the Paleocene. A palynological assemblage of the late Danian *Spinidinium densispinatum* Zone characterizes deposits of the Talitsa Formation. In a wide time range of the *Cerodinium speciosum* Zone, three levels of microflora are distinguished. The lower *Alterbidinium circulum* and middle *Isabelidinium viborgense* levels are established in the late-middle Selandian, and the upper *Cerodinium medcalfii* level is within the late Selandian-early Thanetian interval. As is inferred from analysis of Paleocene palynological assemblages characteristic of marine and continental deposits, the climatic zoning was untypical of the Paleocene, and the West Siberian and Arctic seas have not been connected in the late Talitsa time.

Key words: Paleocene; Talitsa, Tibeisale, and Serov formations; spores, pollen, dinocysts.

INTRODUCTION

Paleocene deposits are exposed or penetrated by prospect boreholes only in the marginal parts of the West Siberian Plate, and additional information is necessary for their paleontological description and stratigraphic analysis. Because of the same situation the northern West Siberian Plate, the Paleocene Talitsa Formation of marine sediments is characterized in the northern part of the Trans-Urals and central lithologic-facies region of the West Siberian Plate only by logging diagrams from deep boreholes. North of the line Yekaterinburg-Tyumen, the Talitsa Formation overlies, with a hiatus, the Upper Cretaceous horizons in most reference sections and boreholes (Akhmet'ev *et al.*, 2001). Dark gray and black dense clay dominates in the lower part of the Talitsa Formation, whereas gray silty clays of normal and opoka-like appearance prevail in the upper part. In the Yamal-Taz region, the facies analog of the Talitsa Formation is the Tibeisale Formation of shallow-water sediments divisible in two subunits. The lower Tibeisale Subformation is represented by dark gray and gray clay with coaly detritus and silt, and the upper one is composed of kaolinite sand with interlayers of coaly siltstone and clay. The Marsyaty and Ivdel formations of the Trans-Urals (Ivdel area) are correlated with the lower and upper Talitsa subformations. The upper Paleocene Serov Formation grades eastward into the lower subformation of the Lyulinvor Formation. Both formations conformably rest here on underlying deposits with a thin basal layer of glauconite sandstone. These lithostratigraphic units are composed of gray to dark gray dense opoka with intercalations of light gray dense clay showing the conchoidal opoka-like fracture. The replacement of opoka by clay is seen in the west-east direction.

On the north and northwest of the West Siberian Plate, spore-pollen assemblages were studied by K.N. Grigor'eva and I.S. Ediger (see in Kulakhmetov, 1965), Chirva and Lyubomirova (1973), and Semochkina *et al.* (1994). Two types of palynological assemblages have been established: the *Trudopollis menneri-Nudopollis endangulatus-Oculopollis giganteus* assemblage of the lower Paleocene and the *Trudopollis menneri-Triatriopollenites aroboratus-Anacolosidites insignis* assemblage of the upper Paleocene. Zonal dinocyst assemblages were established in the northern part of Western Siberia in deposits of the Marsyaty, Talitsa, and Serov formations by Il'ina *et al.* (1994), Vasil'eva (1999), and Yakovleva (2000). They characterize the *Cerodinium striatum* Zone of the Lower Paleocene and two, the *Cerodinium speciosum (C. speciosum-Alterbidinium circulum)* and *Apectodinium homomorphum* zones of the upper Paleocene.

MATERIALS

Paleocene palynological assemblages were studied in 23 samples. The samples have been collected partly in the Yatriya-Lyapin interfluve from deposits of the Talitsa (samples 519/2, 520, and 522) and Serov (samples 554/6, 554, and 552) formations. In addition, undifferentiated Cretaceous-Paleogene deposits (sam-

ples 6 and 8) and the Talitsa Formation have been sampled in borehole sections 21-k, 22-k, and 25-k situated in the Synya, Khulga, Kempazh, and Malaya Ob localities (Fig. 1). The study areas belong to the northern part of the trans-Uralian lithological-facies region of the West Siberian Plate. Generalized results of the spore-pollen analysis of samples from the upper part of the Tibeisale Formation are presented for comparison. The Talitsa Formation is represented most completely in the borehole 25-k section (Fig. 2), where it is 151.5 m thick (depth interval 237.5–86.0 m). Two members are distinguished in this section. The lower member within the depth interval of 237.5–185.0 m is composed of alternating greenish gray to dark brown clay, glauconite sand, and thin interlayers of silt, sand, fine-grained micaceous sandstone, and sandy-clayey silts rich in polymictic gravel and pebble. Inclusions of pyrite, siderite, and organic remains (rare) are typical of these rocks. The upper member (depth interval 185.0–86.0 m) overlies, with a hiatus, the lower member. Greenish gray micaceous clay, sandy in some intervals, is a predominant rock of the upper member. Clay beds are intercalated with small lenses and pockets of silt and silty sand, which show inclusions of pyrite, small brown shell fragments, and fucoids, and with black clay interlayers. At the roof, clay and silt are brown, well graded, with pyrite inclusions and thin laminae of glauconitic sand and sandstone. Seven samples nos. 186 (213.5 m), 184 (212.0 m), 180 (208.0 m), 174 (202.0 m), 169 (197.0 m), 156 (184.5 m), and 106 (120 m) have been collected from these members.

In the borehole section 22-k (166.5–106.5 m), deposits of the Talitsa Formation bear abundant glauconite. The section is composed of greenish black clayey silt with thin interlayers of gray fine-grained sandstone and green thin-laminated clay. At the top, there is greenish silt with interlayers and small lenses of sand. Pyrite is common, but wood fragments are rare in sediments. The formation is 60.0 m thick in this section. The rocks have been sampled at the depth levels of 166.2 m (no. 155), 166.0 m (no. 155), 154.0 m (no. 145), 143.2 m (no. 136), 131.0 m (no. 129), 130.0 m (no. 127), and 119.0 m (no. 118).

In the borehole 21-k, the Talitsa Formation is much thinner (depth interval 150.0–144.0 m) and represented by greenish black homogeneous clay with rare gravely inclusions of quartz. One sample (no. 128) was collected from the depth of 145.0 m. In boreholes 21-k, 22-k, and 25-k, Paleocene deposits overlie with an erosional hiatus the Upper Cretaceous deposits (Berezovo Formation) and are overlain, also with a hiatus, by Neogene rocks. The following rock types of the Talitsa Formation have been analyzed: gray opoka-like siltstone (samples 519/2 and 520), glauconite siltstone (Sample 522), and gray clay (samples 6 and 8). Intervals between sampling levels are equal to 1.0 m. Light gray fine-grained sand (Sample 554/6, 2.0 m) and gray siltstone samples 554 (0.5 m) and 552 (1.3 m) represent rocks of the Serov Formation.

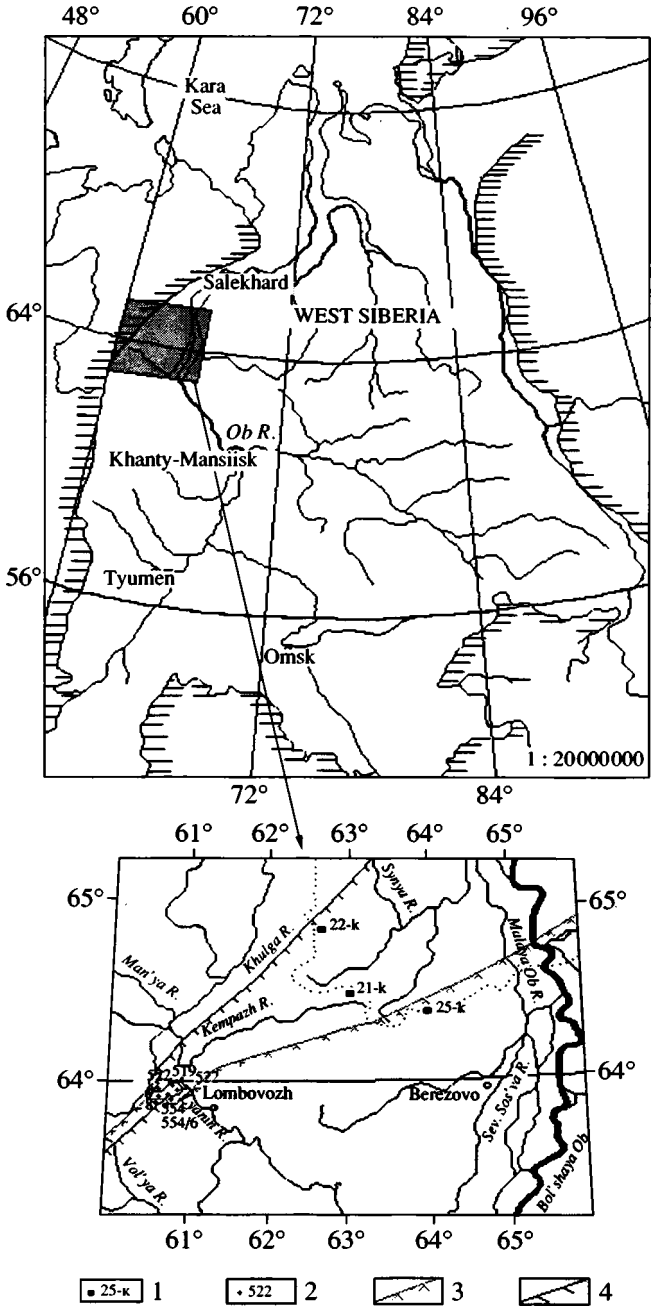


Fig. 1. Locations of studied boreholes and samples: (1) borehole number; (2) sample number; (3, 4) distribution limits of (3) lower and (4) upper Paleocene marine deposits, based on the results obtained.

INVESTIGATION METHODS

The chemical treatment of rock samples suggested by V.P. Grinchuk (*Pyl'tsevoi analiz...*, 1950) and subsequent centrifuging in heavy liquid (specific weight 2.35) were used to obtain macerates from rock samples. The acetolysis was used to get the best separation of organic-walled microphytofossils. Siliceous rocks were treated in hydrofluoric acid twice: (1) after the weigh-

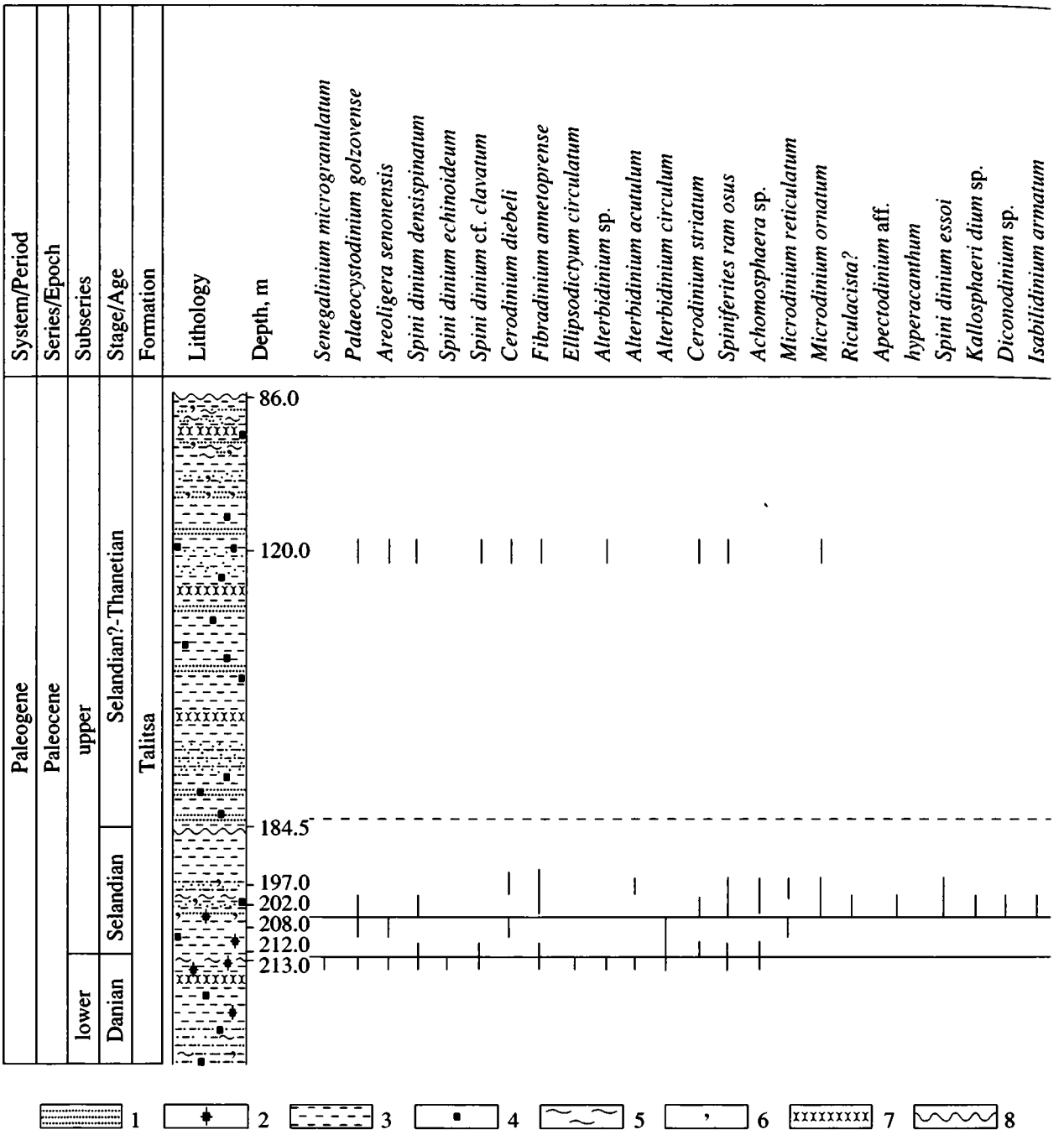
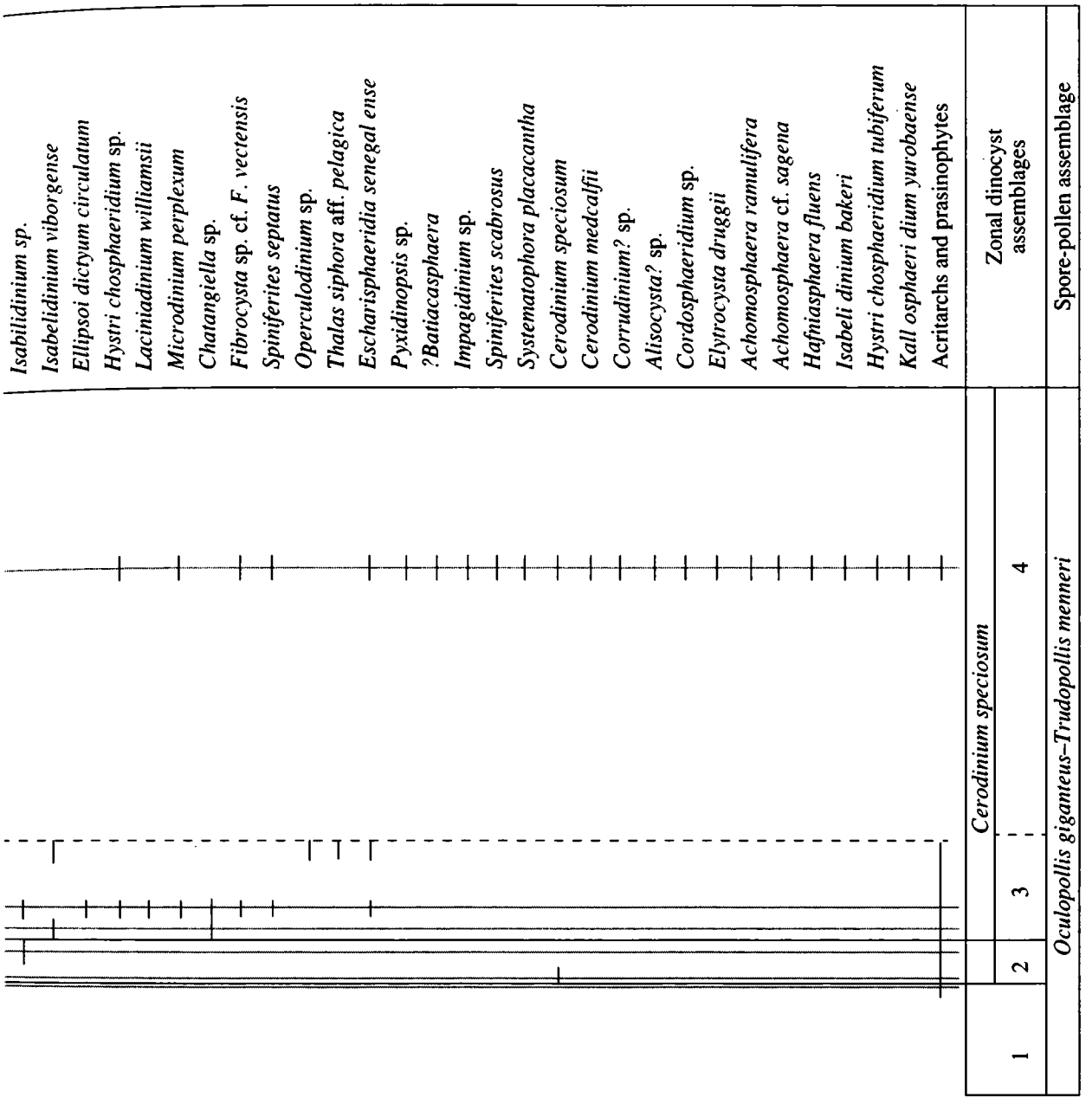


Fig. 2. Distribution of organic-walled microphytoplankton in the Talitsa Formation, Borehole 25-k, the Ob River lower courses: (1) sand; (2) siderite; (3) clay; (4) pyrite; (5) silt; (6) glauconite; (7) sandstone; (8) erosion surface.

ing treatment in 10% alkali solution, but prior to centrifuging in heavy liquid; (2) for the complete decomposition of rocks (Dedovich, 1983). Macerated microphytofossils were identified in temporary and permanent slides under microscopes Biolam R-17 and Laboval 4 using 400- to 600-fold magnification. In permanent slides, macerates are mounted into glycerin-gelatin matrix. Calculation results are subjected to statistical analysis. I calculated percentages of miospores and organic-walled microphytoplankton relative to total amount of specimens in an assemblage. Percentages of each microphytoplankton taxon relative to total amount of palynomorphs and summary contents of miospores in spore-pollen spectra were estimated in addition. For the pollen group, I calculated abundance rates of gymnosperms and angiosperms and percentages of *Normapolles, Postnormapolles, Tricolporopollenites, Castanea-Castanopsis* group, and *Inapertur-*



opollenites (Taxodiaceae) important in terms of stratigraphy and paleoecology. Percentages of dinocysts (Dinophyceae), green algae (Chlorophyceae), acritarchs (*Acritarcha*), and prasinophytes (Prasinophyceae) in the group of organic-walled microphytoplankton (Fig. 3) were used to calculate the offshore-distance coefficients (Akhmet'ev and Zaporozhets, 1996) for their assemblages. By interpretation, I took into consideration the relative abundance (predominant or subordinate) dinocysts versus miospores and the appearance, disappearance, and presence of representative, index, correlative, and zonal species in palynological assemblages. Correlating data on dinocysts from deposits of marine genesis with the Paleogene biozonations of nannoplankton and planktonic foraminifers in the West European paleobasins, I got a chance to establish the formation time of Paleogene sediments in West Siberia based on stage subdivisions of the general stratigraphic scale. Basic in this case were zonal scales elaborated by Heilmann-Clausen (1985, 1994), that of the Paleocene in particular, and by Powel (1992). In addition, I used in calibrations the unified regional stratigraphic scheme of Paleogene deposits in the West Siberian Plain (*Unifitsirovannye...*, 2001).

Five microphytoplankton groups important for paleoecological reconstructions are those of (1) genera *Spiniferites, Areoligera, Cordosphaeridium, Hystrichosphaeridium*, (2) "*Deflandrea*," (3) *Microdinium*,

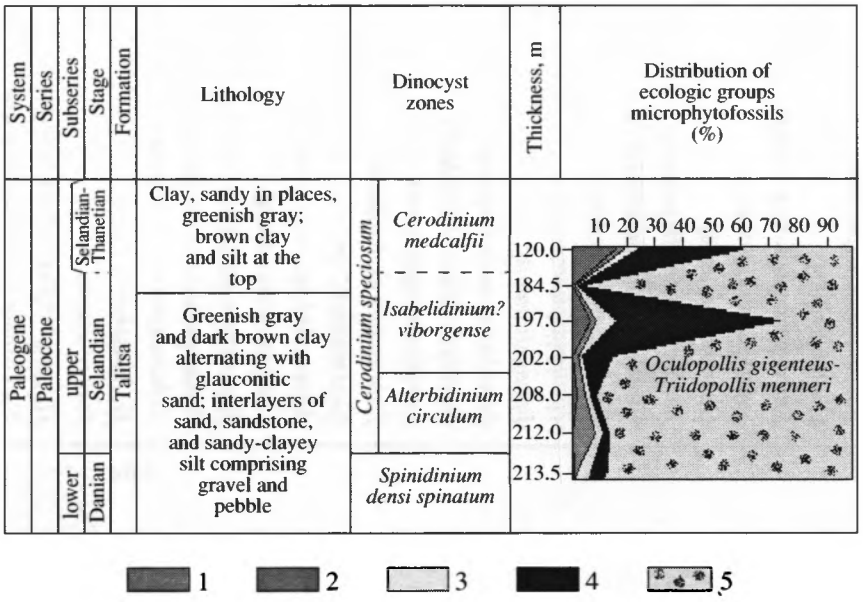


Fig. 3. Distribution of ecologic groups of microphytofossils in the Talitsa Formation, Borehole 25-k, the Ob River lower courses: (1) *Areoligera-Spiniferites-Hystrichosphaeridium-Cordosphaeridium*; (2) "*Deflandrea*"; (3) other dinocysts; (4) acritarchs and prasinophytes; (5) spores and pollen.

and of other (4) dinocysts or (5) acritarchs and prasinophytes. *Spiniferites, Areoligera, Cordosphaeridium* and *Hystrichosphaeridium* species are indicative, when present, of marine setting with normal salinity. Indications of lower salinity and lagoonal (estuarine) sedimentation settings are as follows: prevalence of *Deflandrea* and Wetzelielloidae in the assemblage, a low offshore-distance coefficient, low to moderate dispersal of dinocysts dominated by one or two species, and a mass abundance of acritarchs (Akhmet'ev and Zaporozhets, 1996; Downie *et al.*, 1971; Du Chéne and Adediran, 1984; Köthe, 1990). Forms prevalent in spore-pollen spectra characterize types of plant communities. Presumed phylogenetic relationships between fossil and present-day pollen taxa are useful for discrimination of tropical and subtropical assemblages. Three ecologic groups of palynoflora, which have been distinguished by Gol'bert *et al.* (1977), namely those of hygromesophilic plants, xerophytes, and euryphytes, are regarded in this work as indicative, to some extent, of the climatic parameters and, indirectly, of the paleolandscapes (Manykin, 1973).

RESULTS AND DISCUSSION

All the samples yielded spores, pollen, and algatype sheaths attributed to Acritarcha and Prasinophyceae. Rocks of the Talitsa (except for samples 519/2 and 520) and Serov formations bear dinoflagellate cysts. The Talitsa Formation is most completely characterized by dinocysts and spore-pollen assemblages in the borehole 25-k, where four dinocyst assemblages have been established.

Assemblage 1 (depth level 213.5 m) includes species of the genus *Spinidinium* (*S. densispinatum* Stenley, *S. clavatum* Harland, *S. echinoideum* (Cookson et Eisenack) Lentin et Williams), which associate with *Areoligera senonensis* Lejeune-Carpentier, *Spiniferites ramosus* (Ehrenberg) Loeblich et Loeblich, *Achomosphaera* sp., *Palaeocystodinium golzovense* Alberti, *Fibradinium annetoprense* Morgenroth, *Ellipsoidictyum circulatum* (Clarke et Verdier) Lentin et Williams, *Senegalinium* sp. cf. *S. microgranulatum* (Stenley) Stover et Evitt, *Alterbidinium* sp., *A. acutulum* (Wilson) Lentin et Williams, and *A. circulum* (Heilmann-Clausen) Lentin et Williams. The last form abundant in the second assemblage from a higher level was found as a single specimen. Acritarchs and prasinophytes *Micrhystridium* sp., *Paucilobimorpha apiculata* (Cookson et Eisenack) Prossl, *Verichachium* sp., *Ovoidites* sp., and *Comasphaeridium multispinosum* (Pfstiels) De Coninck are of a low abundance in this assemblage. The content of organic-walled microphytoplankton is 12.9%. Pollen of gymnosperms dominates (54.0%) in the spore-pollen spectrum and is associated with *Inaperturopollenites* sp. (Taxodiaceae), the content of which is as high as 44,1%. *Pinuspollenites* sp., *Piceapollenites* sp., *Podocarpus giganteus* Zakl., *Cedrus* sp., and *Sciadopitys* sp. are found as single specimens. The spore group is represented in the spectrum by Upper Cretaceous relicts *Gleicheniidites* sp., *G. senonicus* (Ross.) Grig., *Gleichenia laeta* Bolch., *Gleichenia delicata* Bolch., *Cicatricisisporites* sp., *Lygodium japoniciforme* E. Iv., *Appendicisporites* sp., *Verrucatosporites* sp., *Laevigatosporites* sp., and *Stereisporites* forms. These forms associate with Paleogene

species *Sphagnum putilum* var. *tenuissimum* Droch. et Purt., *Sh. regium* Drozh. and *Sh. australe* Cook et Pike. Pollen of gymnosperms includes representatives of the *Normapolles* stemma (*Oculopollis giganteus* Zakl., *Trudopollis protrudens* Pflug, *T. nonperfectus* Pflug, and *Simplocacites* sp.) and single specimens of *Orbiculapollis globosus* Clonova, *Anacolosidites insignis* Samoil., *Azonia fabacea* Samoil., *Subtriporopollenites anuloidae* Formen, *Stephanopollenites* sp., and *S. pentaradiatus* Krutzsch. Other taxa of the group are *Tricolporopollenites* cf. *oviformis* (R. Pot.) Kedves, *Triatriopollenites confusus* Zakl., and *Tricolpites microreticulata* type, all described from the Paleocene of West Siberia (Zaklinskaya, 1963) and West Europe (Pflug, 1953; Kunert and Lenk, 1964; Krutzsch and Vanhoorn, 1977). Gymnosperms are abundant throughout the section, except for Sample 106 (120.0 m), in which angiosperms dominate.

Assemblage 2 (depth levels 212.0 and 208.0 m) is of another composition, because species *Cerodinium speciosum* (Alberti) Lentin et Williams, *Cerodinium diebelii* (Alberti) Lentin et Williams, *Isabelidinium* sp., and *Microdinium reticulatum* Vozzhennikova appear here in the dinocyst group. In the sample from the depth of 212.0 m, there was established a predominance of *Alterbidinium circulum* Heilmann-Clausen among algae and an increased abundance of *Areoligera senonensis* Lejeune-Carpentier, while acritarchs were found as single specimens. These changes in the microphytoplankton composition are accompanied by changes in the spore-pollen spectrum, in which pollen of *Anacolosidites insignis* is of an increased abundance and specimens of *Quercites sparsus* (Mart.) Samoil. appear. The dominant taxon of the spore group is *Sphagnum putilum* typical of the upper Paleocene palynological assemblages (Drozhashchikh, 1983). The content of *Inaperturopollenites* spp. (Taxodiaceae) is 32.2 to 44.2%. Pollen of angiosperms is as diverse as in Assemblage 1, though Postnormapolles Pflug is somewhat more abundant and diverse, represented by *Triatriopollenites pseudorurensis* Pfl., *Triatriopollenites myricoides* Kremp., and *Triatriopollenites roboratus* Thomps. et Pflug.

Assemblage 3 (depth levels 202.0, 197.0, and 184.5 m). Samples from depth levels of 202.0 m (the contact between light greenish gray and dark gray clay) and of 184.0 m yielded specimens of the *Isabelidinium ? viborgense* Heilmann-Clausen. The assemblage is lacking *Alterbidinium circulum* but includes other alga species characteristic of the preceding assemblage. Species first appearing in the assemblage are *Microdinium ornatum* Cookson et Eisenack, *Microdinium perplexum* Soncini et Rauscher, *Riculacysta* sp., *Apectodinium* aff. *hyperacanthum* (Cookson et Eisenack) Lentin et Williams, *Kallosphaeridium* sp., *Hystrichosphaeridium* sp., *Fibrocysta* sp. aff. *F. vectensis* (Eaton) Stover et Evitt, *Hafniasphaera septata* (Cookson et Eisenack) Hansen, *Operculodinium* sp., *Thalassiphora pelagica* (Eisenack) Eisenack et Gocht, *Escharisphaeridia senegalensis* Ian du Chéne, *Spinidinium essoi* Cookson et Eisenack, *Diconodinium* sp., *Elipsodictyum circulatum* (Clarke et Verdier) Lentin et Williams, *Isabelidinium* aff. *armatum* (Cookson et Eisenack) Lindgren, *Laciniadinium williamsii* Ioannides, *Chatangiella* sp., and *Isabelidinium* aff. *cooksoni* (Alberti) Lentin et Williams. Some of them are typical of this assemblage only. Other taxa either seem to be redeposited, or transit into higher levels. The latter are *Fibrocysta* sp. aff. *F. vectensis, H. septata, E. senegalensis, M. ornatum, M. perplexum*, and *Hystrichosphaeridium* sp. Among dinocysts, there are identified *Cerodinium striatum, C. diebelii, Spiniferites ramosus, Spinidinium densispinatum, Palaeocystodinium golzovense*, and *Fibrodinium annetoprense*, which appeared earlier. Acritarchs are of the maximum abundance (54.6%) in a sample from the depth level of 197.0 m. The spore-pollen spectrum of the lower sample does not differ from that described above. In the middle one, abundance of spores and pollen is reduced. In sample 156 (depth level 184.5 m) from a bed with magmatic rock pebbles, these components and organic-walled microphytoplankton are less diverse. Microphytoplankton is represented by single specimens of *Isabelidinium ? viborgense, Apectodinium* sp. (deformed), *Thalassiphora* aff. *pelagica, Escharisphaeridia senegalensis*, and *Acritarcha* forms. The spore-pollen spectrum of the assemblage demonstrates prevalence of *Inaperturopollenites* spp. This taxon and *Sphagnum putilum, S. regium*, Postnormapolles Pflug, and Normapolles Pflug characteristic of the assemblage coexist with rare grains of *Tricolporopollenites* sp., *Tricolpites* sp., *Platycariapollis, Caryapollenites*, and *Interpollis suppligensis* (Pfl.) Krutzsch, which are newcomers.

Assemblage 4 (120.0 m) is characterized by high abundance and species diversity of miospores and organic-walled microphytoplankton. In the last group, there are 33% of dinocysts and 36.7% of acritarchs and prasinophytes. Taxa identified in this group are *Cerodinium speciosum, C. striatum, C. diebelii, C. medcalfii* (Stover) Lentin et Williams, *Palaeocystodinium golzovense, Areoligera senonensis, Spinidinium densispinatum, Fibradinium annetoprense, Spiniferites ramosus, S. scabrosus* Clarke et Verdier, *Lingulodinium* sp., *Achomosphaera* aff. *ramulifera* (Deflandre) Evitt, *Achomosphaera* aff. *sagena* Davey et Williams, *Hafniasphaera septata, H. fluens* Hansen, *Microdinium ornatum, M. perplexum, Hystrichosphaeridium* sp., *H. tubiferum* (Ehrenberg) Deflandre, *Fibrocysta* sp. aff. *F. vectensis, E. senegalensis, Isabelidinium bakeri* (Deflandre et Cookson) Lentin et Williams, *Kallosphaeridium yurobaense* du Chéne et Adediran, *Elytrocysta druggii* Stover et Evitt, *Acritarcha* spp., *Fromea laevigata* (Jiabo) Lentin et Williams, *Crassosphaera* sp., *Crassosphaera ? minima* De Coninck, *Cymactiosphaera tortuosa* De Coninck, *Comasphaeridium multispinosum* (Pastiels) De Coninck, and *Membranosphaera* sp. B De Coninck. The summary abundance of species representing the *Areoligera-Spin-*

iferites-Hystrichosphaeridium-Cordosphaeridium group is as high as 16.5%. This is indicative of a large basin depth in the late Selandian-initial Thanetian time, as compared to the deposition time of sediments in the depth interval 213.5–184.5 m. The total content of spores and pollen is up 29.9%, and angiosperms dominate in this assemblage only. The content of gymnosperms is reduced to 11.9%. Angiosperms are mainly represented by pollen of *Normapolles* Pfl. (49.7%) and by associated prevailing species of genera *Oculopollis* (*O. baculotrudens* (Pfl.) Zakl., *O. praedicatus* Weyl. et Krieg., *O. fossulotrudens* (Pfl.) Zakl., *O. giganteus* Zakl.) and *Trudopollis* genera (*T. menneri* Zakl., *T. nonperfectus* Pflug, *Trudopollis retigressus* Weyland et Krieger). According to presence of *Nudopollis* sp., *Sporopollis elaeagnoides* Zakl., *Triatriopollenites rurobituites* (Pfl.) Pflug, *Triporopollenites* sp., *T. robustus* Pflug., *Triatriopollenites* sp., *Tricolpites* sp., *Quercites sparsus, Aquilapollenites* sp., and *Orbiculapollis globosus*, the spore-pollen assemblage is a constituent of the *Oculopollis globosus–Trudopollis menneri* concurrent-range zone. Such a succession in distribution of miospores through the borehole section 25-k characterizes local conditions of sedimentation. Abundant spores of sphagnum and different ferns associate with Taxodiaceae pollen of in the lower part of the section and are indicative of a high humidity, proximity of a vast boggy area, and small basin depth. These environments were especially characteristic of sediments' deposition in the depth interval of 202.0–197.0 m, where abundance of acritarchs and prasinophytes is from 16.3 to 54.6%. Moreover, spore-pollen assemblages from this interval are similar to those of the Tibeisale Formation, both subformations of which are characterized by a high content of spores *Sphagnum* spp. and Taxodiaceae pollen (Kulakhmetov, 1965).

It is difficult to date the assemblages 1 and 2, since the *Spinidinium densispinatum* Zone is not indicated as a characteristic unit of the Paleocene in the Paleogene stratigraphic scheme of the West Siberian plain, and *Alterbidinium circulum* and *Cerodinium speciosum* are zonal species of Zone DN2 spanning a greater interval of the Selandian. Vasil'eva (2001) argued for an older age of the *A. circulum–C. speciosum* assemblage. Sample 186 (depth level 213.5 m) from the lower part of the borehole section 25-k yielded single specimens of *Alterbidinium circulum* and *Spinidinium densispinatum*. The high abundance of *A. circulum* in Sample 184 (depth level 212.0 m) is accompanied by the first appearance of *C. speciosum*. That is why two assemblages are distinguishable within the range of *A. circulum*. Assemblage 1 (depth level 213.5 m) is correlative with Zone Viborg 1 established in the Danian Limestone (Heilmann-Clausen, 1985) and attributed to the late Danian interval of the early Paleocene (Ali *et al.*, 1994; Heilmann-Clausen, 1994) in the range of nannoplankton zones NP3 (part)-NP4 (part). Powell (1992) correlated the *Spinidinium densispinatum* Zone, distinguished between the first appearance of *S. densispinatum* and the first appearance of *C. speciosum*, with Zone Viborg 1 and attributed it to the early Thanetian (Selandian, according to Berggren, 1994) of the late Paleocene. The first appearance of the zonal species in various regions is recorded however from the Danian (*Chronostratigraphy*, 1998; Brinkhuis *et al.*, 1994; Heilmann-Clausen, 1994). The *Cerodinium speciosum* zonal species appears in Denmark at the base of the Viborg 2 Zone of the Kerteminde Marl = the lower zone of the Selandian (Heilmann-Clausen, 1994). Based on these data, I refer the Assemblage 1 to the late Danian of the early Paleocene and the Assemblage 2 to the early Selandian of the late Paleocene.

Assemblage 3 is correlative with Zone Viborg 2 (part) and *Cerodinium speciosum* Biozone (Powell, 1992). It may span a larger stratigraphic range (the Zone Viborg 3 and *Palaeoperidinium pyrophorum* Biozone), but palynological remains are very scarce in the upper part of the assemblage stratigraphic range and above an erosion surface (depth level 184.5 m), and it is difficult to establish here the assemblage upper limitation. Deposits hosting the Assemblage 3 are attributed to the late Paleocene (the middle-late ? Selandian). Occurrence of *Apectodinium* aff. *hyperacanthum* (Cookson et Eisenack) Lentin et Williams is consistent with this inference, as its first appearance in the North Atlantic is detected at the level close to the boundary between the Danian and Selandian (*Chronostratigraphy*, 1998).

Assemblage 4 with *Cerodinium medcalfii* and *Isabelidinium bakerii* is lacking *Isabelidinium ? viborgense*. The *Cerodinium medcalfii* was found (Heilmann-Clausen, 1985; 1994) in Zone Viborg 4 of the Holmehus Formation (late Selandian-early Thanetian). Powell (1992) correlates the Viborg 4 Zone with the *Alisocysta margarita* Zone (Thanetian) of the North Sea. *Cerodinium speciosum, C. diebelii, C. striatum*, and *Fibradinium annetoprense* present in Assemblage 4 suggest that sediments of Sample 106 (depth level 120.0 m) formed not later than in the early Thanetian as the upper boundary of this species range cannot be above the upper limit of Zone NP7 (Powell, 1992). Assemblage 4 is dated to the late Paleocene in the range of the late Selandian-early Thanetian. The late Paleocene age of assemblages 3 and 4 is also evident from the fact that they include a species first identified by Ioannides (1986) as *Fibrocysta* sp. aff. *F. vectensis* (Eaton) Stover et Evitt from upper Paleocene deposits of the Canadian Arctic Archipelago (zones IV and IVa of the Twosnout Creek and South Coast sections).

An assemblage similar to that described from the depth level of 120.0 m of Borehole 25-k is established in deposits of the Talitsa Formation recovered by Borehole 21-k (Sample 128, depth level 145.0 m). However, *Alisocysta* sp. 2 Heilmann-Clausen present in the assemblage from the last locality restricts its age by the early Thanetian. In this assemblage, angiosperms dominate in the spore-pollen spectrum (25.6% of *Nor-*

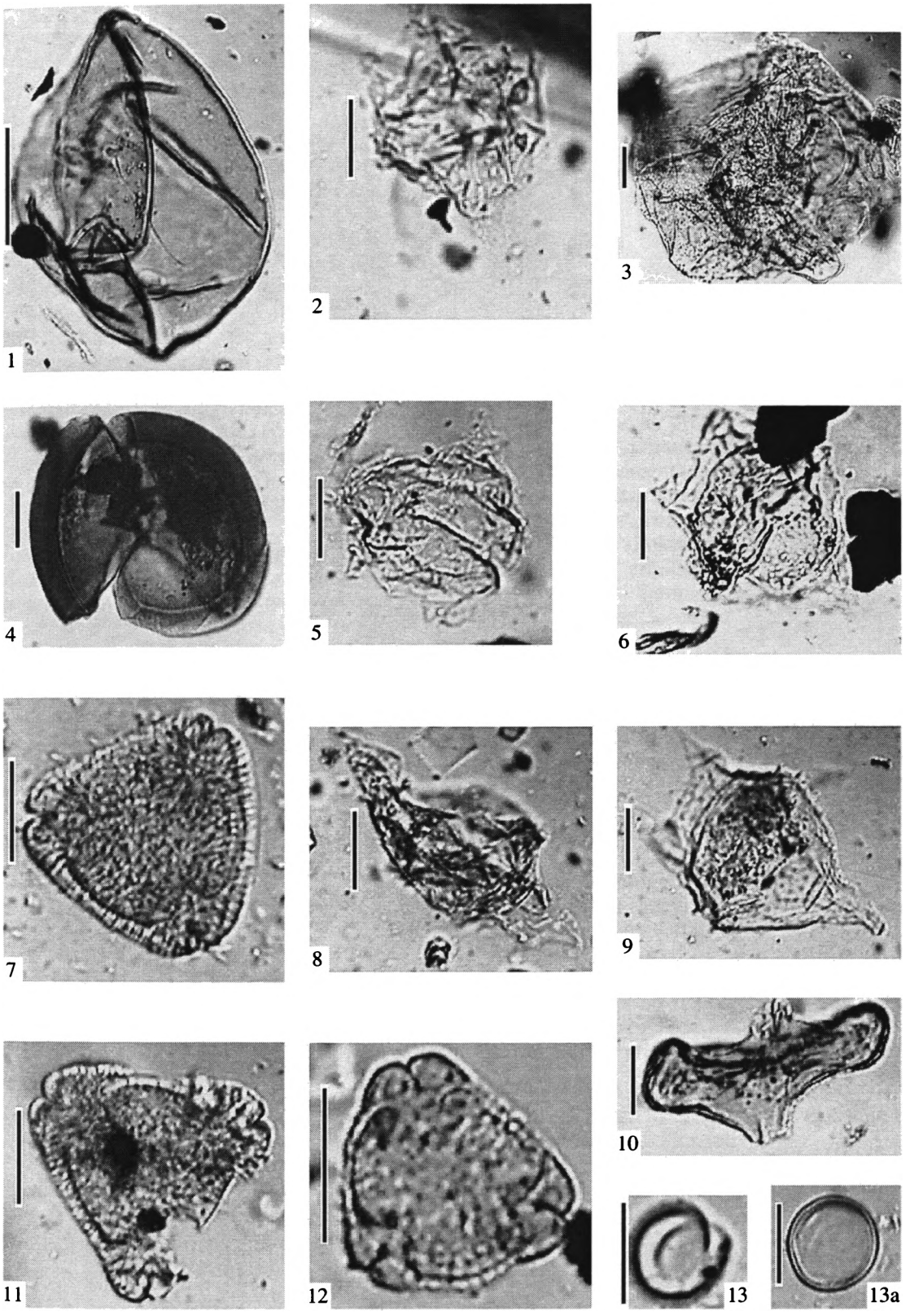
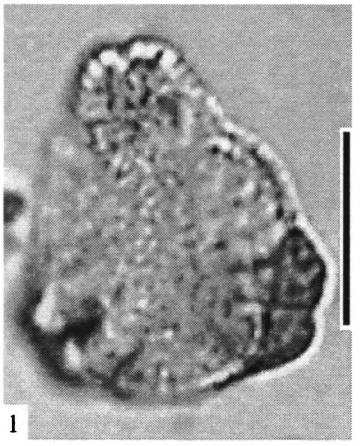


Plate I. Microphytofossils of the Talitsa Formation (Sample 522); scale bar for each image is 25 μm.
(1) *Leiosphaeridia* sp.; (2, 5) *Cerodinium* sp.; (3) *Muratodinium* sp.; (4) *Thytthodiscus* sp.; (6) *Cerodinium medcalfii* Stover; (7, 11) *Oculopollis sibiricus* Zaklinskaja; (8) *Cerodinium diebeli* (Alberti) Vozzhennikova; (9) *C. striatum* (Drugg) Lentin et Williams; (10) *Aquilapollenites* sp.; (12) *Oculopollis* aff. *giganteus* Zaklinskaja; (13, 13a) *Crassosphaera* ? sp.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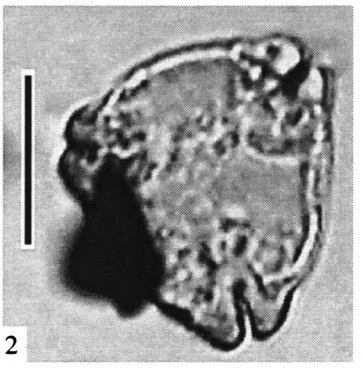
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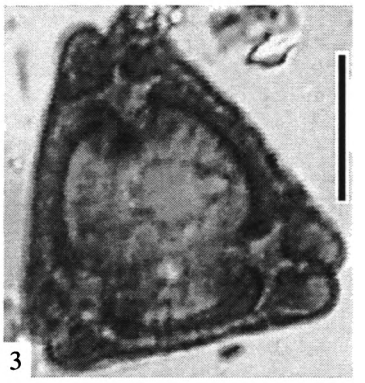
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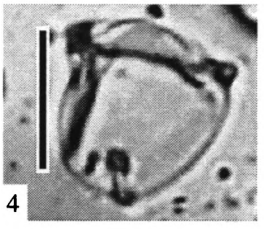
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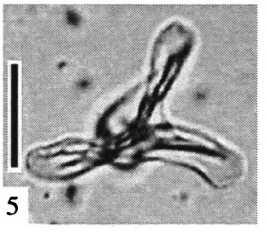
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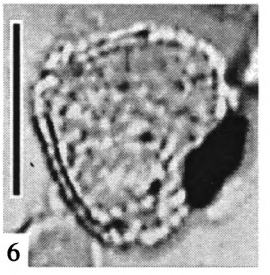
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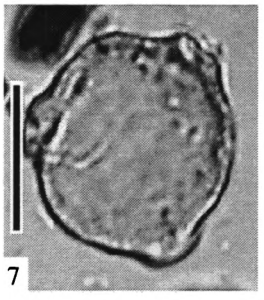
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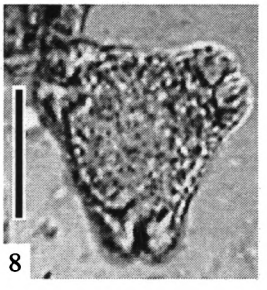
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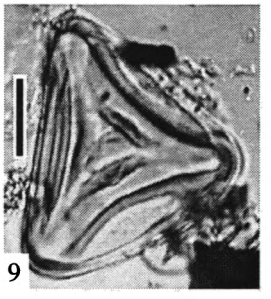
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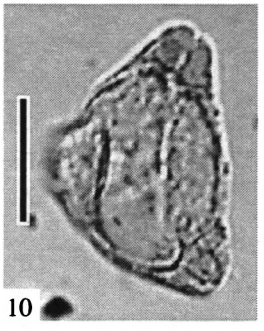
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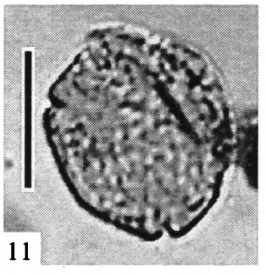
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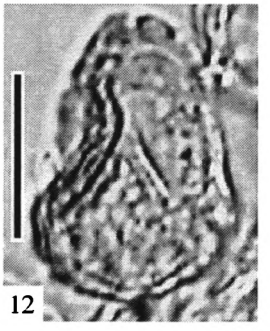
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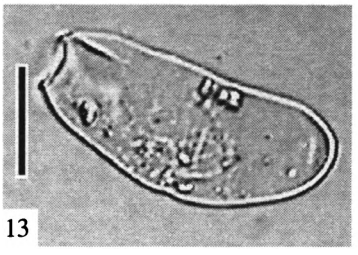
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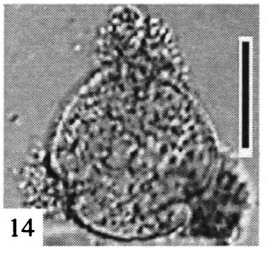
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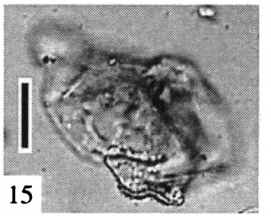
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mapolles pollen only), *Sphagnum putilum* (9.7%) is prevailing taxon of the spore group, and Pinaceae pollen (10.4%) is of prime significance in the group of gymnosperms.

Microphytofossils from the Talitsa Formation rocks recovered by Borehole 22-k are of a uniform composition, and their assemblage is similar to that of Sample 106 (depth level 120.0 m) from Borehole 25-k. Distinctions between some samples are seen in variable proportions between spore, pollen, dinocyst, and acritarch groups rather than in species diversity. This is probably a consequence of changes in the depth and salinity of the basin. The analyzed samples reveal a decrease tendency for the total amount of pollen and an increase of Gleicheniaceae fern spores with a content growth of algae dominated by acritarchs. Organic-walled microphytoplankton of the assemblage consists of *Cerodinium speciosum, Palaeocystodinium golzovense, Spinidinium* aff. *densispinatum, Fibradinium annetoprense, Spiniferites ramosus, S. cornutus* (Gerlach) Stover et Handerbol, *Microdinium reticulatum, Hystrichosphaeridium tubiferum, Fibrocysta* sp. aff. *F. vectensis, Isabelidinium bakeri, Kallosphaeridium* sp., *Thalassiphora inflata* Heilmann-Clausen, *Escharisphaeridia senegalensis, Acritarcha* spp., *Fromea laevigata, F.* aff. *chytra* (Drugg) Stover et Evitt, *Crassosphaera* ? sp., *C.* ? *minima* De Coninck, *Cymactiosphaera tortuosa* De Coninck, *Comasphaeridium multispinosum* (Pastiels) De Coninck, *Membranosphaera* sp. B De Coninck, and of sporadically detected *Pterospermella* sp., *Ovoidites* sp., and *Leiosphaeridia* sp. The host deposits are dated back to the late Selandian-early Thanetian. A great amount of acritarchs may be an indication of sedimentation in lagoons or shoals, and the humid warm climate was favorable for population of ferns in land areas recurrently got free of sea.

Samples 522, 520, and 519/2 from the Talitsa Formation yielded assemblages of low-diverse microphytoplankton (Plates I, II). In Sample 522, I detected *Cerodinium medcalfii* (4.5%), single specimens of *Kallosphaeridium jurobaense, Trithyrodinium* sp., *Phelodinium* ? sp., *Dinocyst* sp., *Comasphaeridium multispinosum, Cimactiosphaera tortuosa, Pterospermella* sp., *Ovoidites* sp., *Fromea laevigata, Baltisphaeridium* sp., *Planctonites* sp., and abundant *Acritarcha* spp. (20.1%). In the dominant angiosperm group (53.9%) of spore-pollen spectrum, prevailing stemma Postnormapolles Pflug (17.3%) contains *Triatriopollenites aroboratus* Pfl., *T. rorubituites* Pfl., *T. roboratus* Pfl., *T. plicoides* Zakl., *T. pseudorurensis* Zakl., *Casuarinidites cainozoicus* Cook. et Pike, *Triporopollenites* sp., *T. robustus* Pfl., and *T.* aff. *giganteus* Pfl. The stemma *Normapolles* (10.5%) consists of *Oculopollis giganteus, O. sibiricus, O. pertinax* Pfl., *Trudopollis nonperfectus, T. retigressus, Pompeckjoidaepollenites subhercinicus* W. Kr., *Basopollis orthobasalis* Pfl., *Nudopollis terminalis* Pfl. subsp. f. sp. *hastaformis* Pfl., and *Extratriporopollenites clarus* Pfl. Besides, few *Anacolosidites insignis* (3.8%), *Quercites sparsus* (3.3%), *Orbiculapollis globosus, Aquilapollenites* sp. were also detected. The Pinaceae pollen (mostly deformed grains) dominates among gymnosperms, whereas Gleicheniaceae prevail in the group of spores. Palynological assemblages from samples 519/2 and 520 are lacking dinocysts, and their acritarchs are of a very low abundance. In the spore-pollen spectrum of Sample 520, the dominant group of gymnosperms (46.2%) is peculiar because of a high content (up to 22.7%) of *Inaperturopollenites* (Taxodiaceae); representatives of *Oculopollis, Aquilapollenites*, and *Orbiculapollis* genera are absent; percentage of *Quercites sparsus* is 5.4%; and *Sphagnum putilum* species are rare. In composition of gymnosperms, the palynological assemblage from Sample 519/2 is similar to that from Sample 522, and in composition of angiosperms, it is analogous to the assemblage from Sample 520. These features represent likely a response to environmental factors. Deposits exemplified by Sample 522 are dated back to the late Paleocene (late Selandian-early Thanetian?). Sediments of samples 520 and 519/2 are likely of the early Thanetian age. Spore-pollen assemblages similar to that of Sample 520 are known from the upper Tibeisale Subformation. This subdivision was penetrated by boreholes 1 (402.5 m), 3 (385.5–170.5 m), 6 (300.0–221.0 m), and 8 (241.5–186.0 m) on the Yamal Peninsula and by boreholes 41 (288.0–287.0 m), 42 (200.0 m), and 49 (284.0–282.0 m) on the Taz Peninsula (the Yamal-Taz lithological-facies region). Sediments of that subformation yield 11.0 to 24.2% of *Stereisporites* (*Sphagnum*) spores, 30.0 to 60.9% of Taxodiaceae pollen, and 45.0% of angiosperms, the representative taxa *Anacolosidites insignis* Samoil. and *Trudopollis menneri* included. Specimens of *Regina excelsa* Samoil. are rare in these sediments, and algae remains have not been found (Semochkina *et al.*, 1994).

Palynological assemblages from samples 6 and 8 (gray clay) of undifferentiated Cretaceous-Paleogene deposits are older, with similar spore-pollen spectra and single alga-like sheaths (*Ovoidites* sp., *Pterospermella* sp., *Cymactiosphaera tortuosa*). In addition, single *Cerodinium striatum* cysts were found in Sample 8. The angiosperm pollen dominant (39.0–63.0%) among

Plate II. Microphytofossils of the Talitsa Formation (Sample 522); scale bar for each image is 25 μm.

(1) *Oculopollis giganteus* Zaklinskaja; (2) *O.* aff. *praedicatus* Weiland et Krieg.; (3) *Trudopollis abnormis* Zaklinskaja; (4) *Orbiculapollis globosus* Chlonova; (5) *Mancicorpus* sp.; (6) *Triatriopollenites* aff. *T. roboratus* Pflug; (7) *Casuariniidites cainosoicus* Cookson et Pike; (8) *Trudopollis nonperfectus* Pflug; (9) *Gleicheniidites senonicus* (Ross.) Grigorjeva; (10) *Trudopollis menneri* Zaklinskaja; (11) *Oculopollis* aff. *fossulotrudens* (Pfl.) Zaklinskaja; (12) *Sporopollis microsulcatus* type (*Simplocas scabripollina* Traverse); (13) *Fromea laevigata* (Jiabo) Lentin et Williams; (14) *Oculopollis sibiricus* Zaklinskaja (deformed); (15) *Proteacidites* sp.

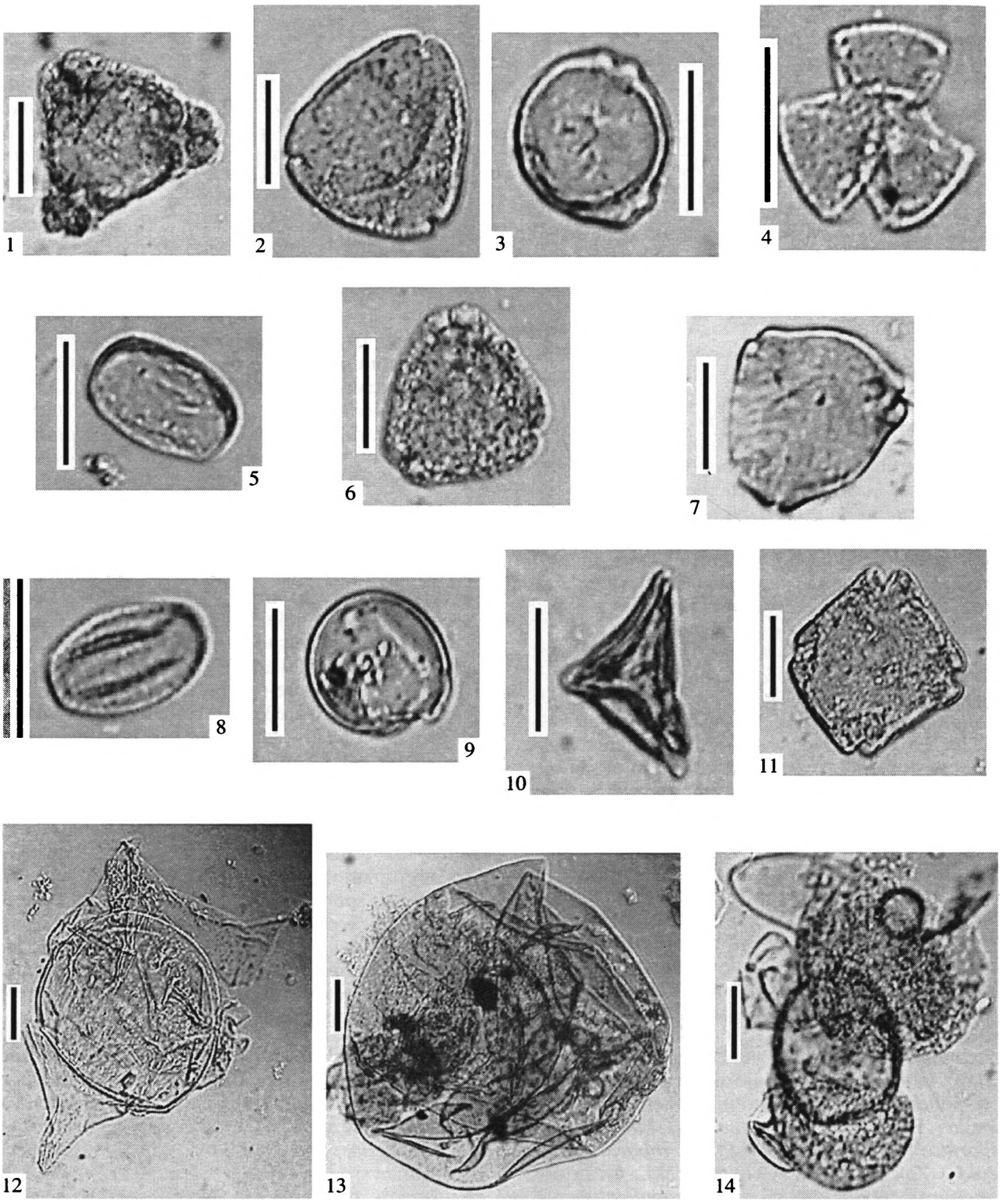

Plate III. Microphytofossils of the Serov Formation (Sample 554/6); scale bar for each image is 25 μm.

(1) *Trudopollis nonperfectus* Pflug; (2) *Triatriopollenites* aff. *roboratus* Pflug; (3) *Triatriopollenites palaeobetuloides* Krutzsch et Vanhoorn; (4) *Tricolporopollenites genuinus* R. Potonie; (5) *Quercites vesca* Agran; (6) *Trudopollis rugosus* (Mart.) Zakl.; (7) *Triatriopollenites myricoides* Kremp.; (8) *Tricolporopollenites gracitricolpatus* Krutzsch et Vanhoorn; (9) *Acritarcha* sp.; (10) *Sporopollis elaeagnoides* Zakl.; (11) *Tetrapollis* aff. *competitor* Weil. et Krieg.; (12) *Deflandrea oebisfeldensis* Alberti; (13) *Leiosphaeridia* sp.; (14) *Podocarpus* sp.

miospores is mainly represented by species of stemma Normapolles Pflug (*Oculopollis giganteus, O. sibiricus, Nudopollis minutus* Zakl., *N. thiergartii* Pfl., *Basopollis* sp., and *Pompeckjoidaepollenites subhercinicus*). The content of gymnosperm pollen ranges from 13 to 34% (Pinaceae gen. sp., *Pinuspollenites* sp., *Piceapollenites* sp., *Cedrus* sp.). Percentage of spores is 24–27%, and they are represented by Late Cretaceous relicts also common in the early Paleocene. These are *Gleichenia* sp., *G. delicata, G. laeta, G. rasilis* Bolch., *Gleicheniidites senonicus* Grig., *Lygodium japoniciforme, Cicatricosisporites* sp., *Appendicisporites* sp., *Selaginella sanguinolenta* (L.) Spring., *Lycopodiumsporites* sp., *Polypodiumsporites* sp., *Stereisporites* sp., *Divisisporites euskirchensis, Leiotriletes gradatus* (Mal.) Bolch., and *Leiotriletes* sp. Their host deposits are likely of the early Paleocene (Danian) age.

Spore-pollen assemblages of the Talitsa Formation from the eastern slope of the Urals were studied by Drozhashchikh (1983) who attributed the *Trudopollis menneri-T. major-Nudopollis tiergartii-Oculopollis solidus* palynological assemblage of the lower Talitsa Subformation to the early Paleocene. According to his data, the *Trudopollis menneri-Triatriopollenites aroboratus-Anacolosidites insignis* palynological assemblage of the upper Talitsa Subformation appears in the section directly below opoka-like clay and never occurs below the base of the foraminiferal *Cibicidoides favorabilis-Gemellides iunatus* Zone. The *Trudopollis menneri-Nudopollis endangulatus-Triatriopollenites aroboratus* (the *Cibicides incognitus* Zone in the lower part of the Talitsa Formation) and *Oculopollis baculotrudens-Extratriporopollenites clarus* assemblages (the Presnov Formation) distinguished in West Siberia are correlated by Panova *et al.* (1990) with the Danian assemblage of the Northern Caucasus.

Samples 554, 554/6, and 552 from the Serov Formation reveal prevalence of gymnosperm pollen (Taxodiaceae and deformed pollen of Pinaceae) in spore-pollen spectra. Angiosperms are represented in these sediments by *Triporopollenites robustus* Pfl., *T. confusus* Zakl., *T. roboratus* Pfl., *T. rorubituites* Pfl., and *T. miricoides*. Percentage of Normapolles pollen (single specimens of *Oculopollis* sp., *O. fossulotrudens* (Pfl.) Zakl., *O. baculotrudens* (Pfl.) Zakl., *Trudopollis* sp., *T. nonperfectus* Pfl., *T. menneri*, and *Pompeckjoidaepollenites subhercinicus*) is not greater than 6.5%, or this pollen is missing. Single grains of *Interpollis suppligensis* (Pfl.) Krutzsch, *Quercus* sp., *Q. vesca* Agran., and *Casuarinidites cainozoicus* Cook. et Pike are detected in the rocks. *Tricolpites* sp., *Tricolporopollenites* sp., *Platycaryapollis* sp., *Caryapollenites* sp., *Ulmoideipites tricostatus* Andersen, *Platanus* sp., and *Polyporites* sp. are of sporadic occurrence. Palynological assemblage of Sample 554/6 is most representative, containing up to 41.5% of algae sheaths (Plate III). The latter correspond to *Cerodinium speciosum* subsp. *glabrum* (Gocht) Lentin et Williams, *Deflandrea oebisfeldensis* Alberti, *Batiacasphaera* sp., *Operculodinium* sp., few *Leiosphaeridia* sp., *Crassosphaera* sp., *Pterospermella* sp., *Fromea laevigata* (19.6%), and to forms identified as *Acritarcha* spp. (17.2%). Species *Paucilobimorpha triradiata* De Coninck (0.3–0.9%) appears at this level. Samples 554 and 552 barren of dinocysts yielded respectively 9.6 and 7.2% of acritarchs. Sediments of the Serov Formation were likely deposited in lagoonal (Sample 554/6) or boggy areas (samples 554 and 522). *Cerodinium speciosum* subsp. *glabrum* and *Deflandrea oebisfeldensis* present in the group of organic-walled microphytoplankton suggest that it is correlative with the *Apectodinium homomorphum-Cerodinium speciosum* subsp. *glabrum* assemblage known in the northern part of the Turgai region and in southern areas of the trans-Urals (Vasil'eva *et al.*, 2001). Further northward (central trans-Urals), opoka beds of the Serov Formation have not been studied in detail with respect to dinocysts. The spore-pollen assemblage of samples 554, 554/6, and 552 is typical of the late Paleocene, and I attribute the corresponding sediments to the Thanetian Age.

CONCLUSIONS

Palynological assemblages studied in the northern part of the trans-Urals within the lithologic-facies region of the West Siberian plain appeared to be correlative, based on representative and zonal dinocyst species, with the *Cerodinium striatum* and *Spinidinium densispinatum* zones of the lower Paleocene and with the *Cerodinium speciosum* and *Cerodinium speciosum glabrum-Apectodinium homomorphum* zones of the upper Paleocene (the latter correlation is arbitrary, mainly based on spores and pollen). Three levels distinguishable in the *C. speciosum* Zone can be regarded as subzones named after representative species. The lower *Alterbidinium circulum* Subzone corresponds to the early Selandian and is distinguished because of first appearance of *C. speciosum* and a sharp increase in abundance of *A. circulum*. The middle *Isabelidinium ? viborgense* Subzone is attributed to the middle Selandian and spans the range of index species. The upper *Cerodinium medcalfii-Isabelidinium bakeri* Subzone is dated back to the late Selandian-early Thanetian based on appearance of its index species and associated *Fibradinium annetoprense* and *C. speciosum*. Judging from the composition of the palynological assemblages, sediments of the Talitsa and Serov formations were deposited mostly in shallow-water marine or lagoonal basins with a low water salinity. The basin attained the largest depth during the late Selandian-early Thanetian period of sedimentation (Sample 106, depth level 120.0 m in Borehole 25). The continental genesis of the Tibeisale Formation deposits suggests that West Siberian and Arctic seas have not been connected in the late Talitsa time. The correlation of palynoflora and organic-walled algal flora from the studied sections leads to the conclusion that the

angiosperm pollen with prevailing Normapolles forms has the peak abundance at the level of the *C. striatum* Biozone (early Paleocene). At that time, land areas with excessive moisture around a sea basin were populated by plant communities producing the Normapolles pollen. The development of Postnormapolles and *Inaperturopollenites* (Taxodiaceae) started in the late Danian, i.e. from the level of *S. densispinatum*, and the angiosperm/gymnosperm (mostly Taxodiaceae) abundance ratio might be determined in the late Paleocene by environmental factors. A great abundance of acritarchs, Taxodiaceae pollen, and deformed Pinaceae pollen in palynological assemblages is indicative of the proximity of a vast boggy area, since present-day Taxodiaceae inhabit bogs, riverbank plains, and sea coasts. Moreover, the Pinaceae pollen of assemblages might be related in origin with a highland of the region, since representatives of Pinaceae grow only in mountains of recent tropical areas. The climate of the Paleocene presumably was subtropical and humid; it was closer to tropical (paratropical) climate in the early Paleocene. The main type of vegetation corresponded to coniferous-deciduous or deciduous-coniferous forests, boggy in places. As the studied palynomorphs appear to be correlative with coeval palynological assemblages of remote regions (Crimea-Caucasus region, northern West Europe), it is possible to assume that climatic zoning was either weak or not manifested at all in the Paleocene.

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Position of the Pontian of the Eastern Paratethys in the Magnetochronological Scale

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Abstract—A comprehensive analysis of paleomagnetic records, nannofossils, mammal fauna, and fission-track dating results was conducted for the Upper Miocene sediments of the Eastern Paratethys. The age of Pontian boundaries in the Eastern Paratethys (Euxinian basin) is determined. The stage base is at 7.5 Ma, the top at 6.7–6.6 Ma, and the lower-upper Pontian boundary at 7–7.1 Ma. In the magnetochronological scale, the Pontian corresponds to the uppermost Chron 7 (C4n) coupled with a greater portion of Chron 6 (C3Br–C3Ar). The lower Pontian is correlated with the upper part of the Tortonian and with an upper part of Zone MN12. The upper boundary of the lower Pontian coincides with the Tortonian-Messinian boundary. The upper Pontian corresponds to the lower third of the Messinian and to the lower part of Zone MN13. The Euxinian upper Pontian is correlated in the Caspian basin with the lower part of the Shemakhan regional substage of the middle Pontian in Azerbaijan. The Zone NN11 corresponds to the upper Maeotian, whole Pontian, and the lowermost Azov Beds of the Kimmerian.

Key words: stratigraphy, correlation, Miocene, Pontian, Eastern Paratethys, Mediterranean, magnetochronology, nannoplankton.

INTRODUCTION

There is no agreement among the geologists who studied during several decades the Pontian of the Eastern Paratethys, regarding its duration, boundary ages, and correlation with the units of the Mediterranean zonation, as exemplified even by the latest works (Steininger, 1999; Zubakov, 2000; Popov and Nevesskaya, 2000; Chumakov, 2000; Snell *et al.*, 2000, 2001; Popov *et al.*, 2001; and others) (Fig. 1).

In this paper, we consider once again the Pontian position in the magnetochronological scale and its correlation with the Mediterranean stages. Over 20 years have passed since our publication (Semenenko and Pevzner, 1979) devoted to this problem. The formerly unknown data became available during this time. New nannofossils were found in the Maeotian, Pontian, and Kimmerian sediments (Bogdanovich and Ivanova, 1997; Lyul'eva, 1989; Semenenko *et al.*, 1999). Results of fission-track dating were obtained for the Pontian and Maeotian deposits in the Black Sea (Euxinian) and Caspian basins (Chumakov *et al.*, 1992). New localities of mammal remains in the Sarmatian, Maeotian, and Pontian deposits were discovered and the formerly known sites were revised. A series of nannofossil datums was verified. Finally, ages of boundaries between magnetochronological units are now considered to be older (Cande and Kent, 1995). The last inference substantially changes former ages of boundaries of stratigraphic units under discussion and, in particular, the Pontian-Kimmerian boundary age in the Zheleznyi Rog Cape section. Nevertheless, the previous conclusion that the Black Sea Pontian is of a "short" time span and correlative with the uppermost Chron 7 and greatest interval of Chron 6, remains unchanged despite the all new data.

Let us remind that the former inference was based on the magneto- and biostratigraphic records in the Zheleznyi Rog and the Chegerchin Trough sections recovered by three boreholes at the site located eight kilometers southeast of the Kazantip Cape, near the Azovskoe Village, Kerchenskii Peninsula (Pevzner and Chikovani, 1978; Semenenko and Pevzner, 1979; *Stratigrafiya SSSR*, 1986, p. 138). Nannofossils found in samples, which have been collected by Pevzner for the paleomagnetic analysis from borehole sections of the Chegerchin trough (Semenenko and Lyul'eva, 1978) and Zheleznyi Rog Cape (Semenenko and Lyul'eva, 1982), were of great importance for interpretation of magnetostratigraphic data. When determining at that time the Pontian position in the magnetochronological scale, we assumed that this stage is represented in its full range in the Zheleznyi Rog Cape section. Since the Pontian sediments in this section show mainly a reversed polarity, they can correspond only to a single paleomagnetic chron (or to a part of it). This implies that the Pontian time span cannot be long. For-

† Deceased.

merly, some authors estimated the Pontian range to be from 3 to 4 m.y. long and dated the base and top of the stage at 9.5–8 and 6–5.5 Ma, respectively. Within this time span, the Pontian sediments with reversed polarity could be correlated with Chron 6 or with the lowermost Gilbert Chron, if one would interpret the paleomagnetic records only.

It should be noted that many researchers correlate the Pontian with the lowermost Gilbert Chron and the Maeotian with Chron 5 and 6 (Trubikhin, 1984; Trubikhin *et al.*, 1984; Svetlitskaya, 1995; Chepalyga and Svetlitskaya, 1995; Snell *et al.*, 2000, 2001; Popov and Nevesskaya, 2000; Popov, 2001; Popov *et al.*, 2001; and others). This interpretation means a significant rejuvenation of the Maeotian and Sarmatian stages and contradicts the radiologic dates obtained by the K/Ar and fission-track methods.

Before discussing the new records, we should emphasize that the Pontian of the Eastern Paratethys is considered in this work as the Pontian of the Black Sea basin. The Pontian regional stage of the Eastern Paratethys is divided in two substages: the lower Novorossian Substage with the Evpatoriya and Odessa beds, and the upper substage with the Portaferian and Bosphorian beds (*Stratigrafiya SSSR*, 1986). After the early Pontian time, an integral former basin was separated into the Black Sea and Caspian isolated basins.

In the Caspian basin, the second half of the Pontian is of a different range and evolution history. In this area, the Pontian is divided into three units: the lower Novorossian, middle Shemakhan, and upper Babadzhan regional substages (*Geologiya SSSR*, 1972). Fauna of the lower substage corresponds to that of the lower Pontian in the Euxinian basin. The middle substage and the upper Pontian of the Euxinian basin bear the molluscan forms in common, while the upper substage is characterized by a very peculiar fauna and can be correlated with the lowermost Kimmerian of the Black Sea region (Andrusov, 1917).

MAGNETOCHRONOLOGICAL SCALE CK95

The magnetochronological scale CK95 (Cande and Kent, 1995) used to determine ages of the unit boundaries is calibrated now based on the astrochronological records for the last 7 Ma instead of previously accepted K/Ar dates, and the Cretaceous-Paleogene boundary is dated therewith at 65 Ma instead of 66 Ma. Accordingly, boundaries between paleomagnetic units became older in the upper part of the scale and rejuvenated in the lower part, as compared to the previous magnetochronological scales.

For the last 13 million years, ages of boundaries between paleomagnetic units in the magnetochronological scale CK95 are older than the analogous boundaries in the previous scales. For the last nine million years, they are older approximately by 10%. Ages of plankton datums and stratigraphic boundaries are now

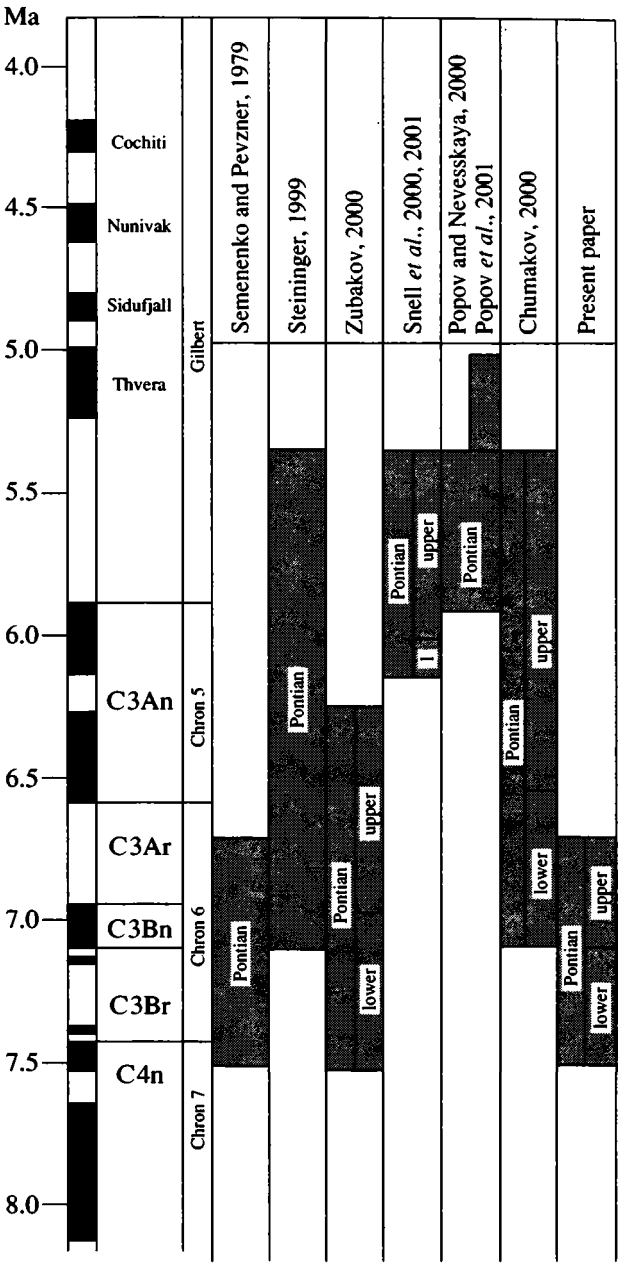

Fig. 1. Magnetochronological scale of the Eastern Paratethys and time spans of the Pontian according to various authors.

determined according to the CK95 scale. To correlate sediments dated by K/Ar (or by fission-track dates calibrated against the K/Ar ages) with that scale, it is necessary to increase their ages by 10%. Without due regard for this fact, the results of correlation would be invalid.

Let us consider, as an example, the data on correlation between lower boundaries of the Pontian in the Eastern Paratethys and of the Messinian in the Mediterranean region (Steininger, 1999; Chumakov, 2000). The Messinian lower boundary (7.1 Ma) is correlative

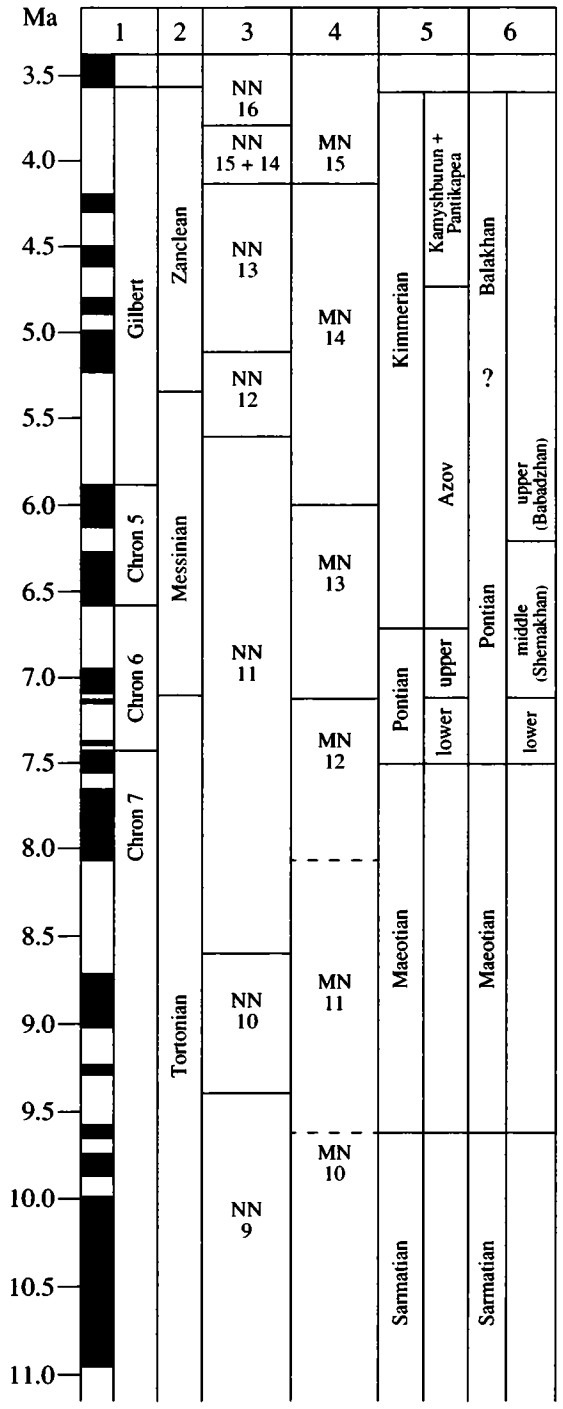

Fig. 2. Correlation of regional stages of the Eastern Paratethys with the magnetochronological scale, Mediterranean stages, nannofossil and mammal zones: (1) magnetochronological scale (CK95); (2) Mediterranean stages; (3) nannofossil zones; (4) mammal zones; (5, 6) regional stages and substages of the (5) Euxinian and (6) Caspian basins.

with the Pontian base dated by fission-track method at 7 Ma (Chumakov *et al.*, 1992). However, such a correlation is incompatible with the paleomagnetic records. In the magnetochronological scale, the base of the Messinian is placed in the middle of Chron 6, and the uppermost Tortonian sediments show a reversed polarity (Fig. 2). In the Eastern Paratethys, a thick zone of the upper Maeotian sediments with normal polarity is situated just below the Pontian deposits (Fig. 3). Regarding the Pontian base as correlative with the Messinian lower boundary, we come to an ambiguous conclusion that the upper part of the Tortonian referred to the lower half of Chron 6 is correlative with the normally magnetized sediments of the upper Maeotian. Increasing fission-track dates for the Pontian rocks by 10%, we avoid this contradiction. For instance, the fission-track date of 6.4 Ma obtained for the lower Pontian top (Chumakov *et al.*, 1992) will be equal to 7 Ma, if increased by 10%, and thus almost identical to the age of the Messinian base. The fission-track date 7 Ma estimated for the Pontian base will be as high as 7.7 Ma after the 10% increase. In this case, we also eliminate the contradiction between radiologic and paleomagnetic data, because the lower Pontian will correspond now to the lower half of Chron 6 and to the uppermost Chron 7.

FISSION-TRACK AGES OF SEDIMENTS IN THE EASTERN PARATETHYS

It should be noted that the fission-track dates obtained for the lower Pontian sediments and for the base of the Shemakhan substage (Chumakov *et al.*, 1992) confirmed not only the correctness of correlation of the Pontian with Chron 6, but also the ages of the lower Pontian boundaries established based on paleomagnetic data (Semenenko and Pevzner, 1979).

All the fission-track dates published by Chumakov were obtained using the decay constant of the ^{238}U—$\lambda_f = 7.03 \times 10^{-17}$ yrs^{-1}. As noted by Kashkarov *et al.* (1987, p. 20), "Special dating of volcanic tuff with the known K/Ar age (tuff sample from the Fish Canyon, Colorado) in many laboratories of the world indicates that the most compatible results can be obtained using the decay constant $\lambda_f = (7.03 + 0.11) \times 10^{-17}$ yrs^{-1} (Roberts *et al.*, 1968). This value was subsequently recommended for a fission-track dating (Naeser *et al.*, 1981)..."

To avoid a revision of their standpoint, proponents of a higher Pontian position in the magnetochronological scale recalculated the fission-track dates from works by Chumakov using the decay constant $\lambda_f = (8.24) \times 10^{-17}$ yrs^{-1} (Golovina *et al.*, 1989). Their results significantly rejuvenate ages of stratigraphic boundaries in the Eastern Paratethys. For instance, the Pontian base age of 7 Ma was recalculated into 5.84 Ma, and the last value was regarded as the evidence in favor of referring the Pontian sediments with reversed polarity to the lowermost Gilbert Chron. It is obvious, however, that such a manipulation with the fission-track dates by using the different decay constant of ^{238}U is incorrect. This manipulation lacks any physical substantiation, and the

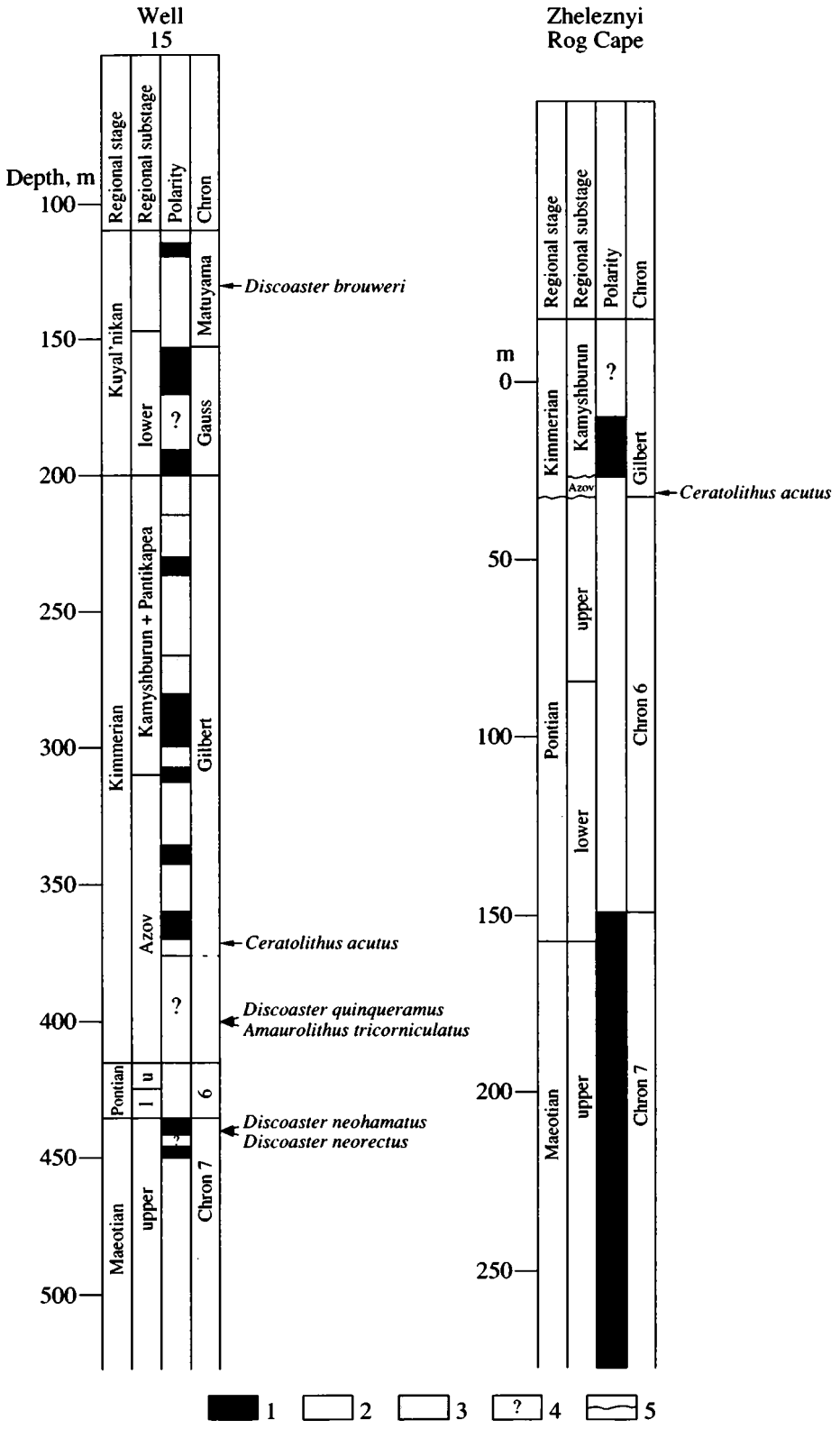

Fig. 3. Paleomagnetic characteristics of Well 15 and Zheleznyi Rog Cape sections and position of nannofossil findings: (1) normal, (2) reversed, and (3) anomalous polarity; (4) break in paleomagnetic records; (5) stratigraphic hiatus.

Age (Ma) of boundaries between stratigraphic units of the Eastern Paratethys

Regional stage	Regional substage	Nanno-fossils	Mam-mals	Fission-track dates	Paleomag-netic records
Kimmerian	upper				3.6
	middle				4.7
	lower	>5.6	>6.57		6.7
Pontian	upper		7.1	7.0	
	lower	>7.4		7.7	7.5
Maeotian		<8.7 >9.4		<10.2	9.6
		<10.7			

ages recalculated for Pontian sediments of the Eastern Paratethys cannot be used for stratigraphic purposes.

The dates in question (Chumakov *et al.*, 1992) were obtained for the Pontian beds of volcanic ash in the Azerbaijan sections (Caspian basin). Thus, only the age determinations for the lower boundary of the Novorossian regional Substage and for the base of the Shemakhan substage (= top of the lower Pontian) are applicable to the Pontian of the Euxinian basin. There is a number of consecutive dates for the Shemakhan regional substage. Its top is as old as 5.6 Ma, or 6.2 Ma after the increased by 10%. There was a single age measurement, 5.19 ± 0.89 Ma, obtained for the Babadzhan Substage, and it is inappropriate even for approximate evaluation of the unit age owing to a large analytical error and unclear stratigraphic position of the analyzed sample (Chumakov *et al.*, 1992, p. 51).

Thus, the fission-track ages (Chumakov *et al.*, 1992) increased by 10% lead to the following inferences: the Maeotian base is not older than 10.2 Ma, the Pontian base is as old as 7.7 Ma, and the top of the lower Pontian is 7.0 Ma old with an accuracy of ±8% (table).

NANNOFOSSILS FROM THE MAEOTIAN, PONTIAN, AND KIMMERIAN DEPOSITS OF THE EASTERN PARATETHYS

The Maeotian is subdivided in two regional substages, the lower Bagerovian and the upper Akmanaian (*Stratigrafiya SSSR*, 1986). In the Maeotian sediments of the Crimea, Well 501 on the Akmanaiskii Isthmus of the Kerchenskii Peninsula, the *D. hamatus* Zone NN9 was recognized in the lower part of the Bagerovian Substage that yields *Discoaster hamatus* Mart. et Bram. and *Catinaster calyculus* Mart. et Bram. The upper part of the Bagerovian and the lower of the Akmanaian substages are correlated with the *D. calcaris* Zone NN10 (= *Discoaster neohamatus* CN8 and the *Discoaster bellus* Subzone CN8a; Bogdanovich and Ivanova, 1997). Therefore, the boundary between zones NN9/NN10 is recorded in the lower Maeotian sediments. Its age is 9.4 Ma (Berggren *et al.*, 1995), indicating that the Maeotian base is older than this date. Species *Catinaster calyculus* Mart. et Bram. present in the lowermost Maeotian restricts the stage basal age by the appearance time of that species at 10.7 Ma (refer to the table). In Well 15 near the Kazantip Cape, Kerchenskii Peninsula (Fig. 3), the uppermost Maeotian sediments yield *Discoaster neorectus* Bukry (Semenenko and Lyul'eva, 1978; Semenenko and Pevzner, 1979). Stratigraphic ranges of *Discoaster neorectus* Bukry and *Discoaster loeblichii* Bukry completely coincide (Perch-Nielsen, 1985). The latter appeared at the level of 8.7 Ma and became extinct at 7.4 Ma (Berggren *et al.*, 1995). Thus, *Discoaster neorectus* found in the uppermost Maeotian interval of Well 15 indicates that the Maeotian-Pontian boundary is not younger than 7.4 Ma and no older than 8.7 Ma.

Nannofossils of the lower Pontian sediments are known from several sections of the Eastern Paratethys, but the zonal species are missing there. The index species of the *Discoaster quinqueramus* Gart. Zone NN11 with the LAD at 5.6 Ma (Berggren *et al.*, 1995) was encountered in the lower Azov strata of the Kimmerian, Well 15 in the Chegerchin Trough (Semenenko and Lyul'eva, 1978; Semenenko and Pevzner, 1979). This finding indicates that the Pontian top is older than 5.6 Ma. *Discoaster quinqueramus* present in the Kimmerian deposits excludes a possibility that the Bosphorian beds of the upper Pontian is correlative with the *A. tricorniculatus* Zone NN12, as was suggested by Snell *et al.*, (2000, 2001).

In Well 15, *Ceratolithus acutus* Gartner et Bukry was found in sediments with reversed polarity of the Azov regional substage of the Kimmerian immediately below the normal polarity zone that is correlated with the Thvera Subchron (Semenenko and Pevzner, 1979). Semenenko found this species in the Zheleznyi Rog Cape section also in deposits with the reversed polarity of the Azov Substage (see below). The occurrence of *Ceratolithus acutus* in the apparently lower Kimmerian sediments of the Euxinian basin indicates that their range corresponds to the upper part of the Zone NN12 (= *C. acutus* Zone CN10b).

MAMMAL FAUNA OF THE PONTIAN

Researchers who study the Pontian mammal fauna of the Eastern Paratethys have controversial opinions with respect to correlation between these faunas and MN zones. Gabuniya (1986) assigned the lower Pontian mammals to Zone MN12, Korotkevich (1988) to the upper part of Zone MN13, Krakhmalnaya (1996) and Topachevskii *et al.* (1997, 1998) to Zone MN14.

Mammal remains associated with marine sediments of the upper Pontian are not yet found on the CIS territory. The mammal fauna from the upper Pontian deposits with reversed polarity (above the beds bearing *Congeria rhomboidea* M.Hoern.) is known from the Pannonian Basin, the Hatvan locality of Hungary (Pevzner, 1986), where it is referred to the lower half of Zone MN13 (Mein, 1990).

The lower Pontian sediments of Odessa (station 16 of Bol'shoi Fontan) and Vinogradovka localities yield remains of *Schizogalerix* and *Spermophilinus* (Topachevskii *et al.*, 1997, 1998), which indicate that the localities are not younger than Zone MN13, since the LAD of both genera is within this zone (Mein, 1999). *Prolagus crusafonti* Lopez is present in both localities. In the middle of Zone MN13, this species is replaced by *P. michauxi* Lopez (Lopez Martinez, 1977). Thus, the fauna from these localities cannot be referred to the upper half of Zone MN13 and especially to Zone MN14. *Tetralophodon longirostris* (Kaup) found in the lower Pontian limestones near Odessa indicates, in opinion of Gabuniya (1986), that the lower Pontian mammal fauna is correlative with Zone MN12, as extinction of this genus is recorded in the uppermost part of the latter (de Bruijn *et al.*, 1992). The composition of small mammal fauna from the lower Pontian deposits of Odessa and Vinogradovka sites does not contradict this conclusion. Thus, the lower Pontian mammal fauna is referred to Zone MN12 and that of the upper Pontian to the lower half of Zone MN13 (Fig. 2).

The upper boundary of Zone MN13 is dated at about 6 Ma and corresponds to the uppermost Chron 5 (Pevzner *et al.*, 1996, 2001). The zone lower boundary is ca. 7.1 Ma old and is correlated with the base of the Messinian and with the beginning of the global C-isotope excursion (Bernor *et al.*, 1996; Steininger *et al.*, 1996).

As it follows from data considered above, the Pontian upper boundary is apparently older than 6 Ma. Taking into account that sediments with reversed polarity correspond to Chron 6 at the Hatvan locality, one can suggest that this boundary is older than the upper limit of Chron 6 dated at 6.57 Ma. The lower-upper Pontian boundary is close in age to the base of Zone MN13 that is 7.1 Ma old (table).

INTERPRETATION OF PALEOMAGNETIC RECORDS IN THE ZHELEZNYI ROG CAPE SECTION

The upper Maeotian, Pontian, and Kimmerian sediments are exposed in the Zheleznyi Rog Cape section of the Tamanskii Peninsula. The Maeotian sequence over 200 m thick is represented here by light gray thin-bedded clays in the lower and upper parts of the section, and by dark gray, commonly bituminous clays in the middle part. The deposits are almost barren of molluscan fauna. Only in a single interval, 150 m below the base of the Pontian, Semenenko identified *Abra tellinoides* Sinz., *Congeria panticapaea* Andrus., *Paphia abichi* (R. Hoern.), and *Hydrobia* sp., which confirm the Maeotian age of the studied sequence. The Maeotian-Pontian boundary is rather confidently placed at the contact between the lower 3-m-thick member of light gray, almost white clay and the overlying member of dark gray bituminous clays. Small flattened thin-walled and thin-ribbed cardiid shells were found at the base of the latter, whereas *Paradacna abichi* (R. Hoern.) typical of the Pontian was found 2.5 m above.

The Pontian deposits are represented by a 123-m-thick sequence of light gray to bluish thin-bedded clays intercalated with rare beds of darker, locally bituminous clay. *Dreissena rostriformis* Desh., *D. anisoconcha* Andrus., *Paradacna abichi* (R.Hoern.), *Didacna planicostata* (Desh.), *D. paucicostata* Desh., *D. incerta* Desh., *Plagiodacna carinata* (Desh.), *Valenciennesia* sp., and others (identified by Chikovani) are molluscan species common of the Pontian sediments. *Congeria subrhomboidea* Andrus. that marks the boundary between the lower and upper Pontian, was encountered at the level of 73 m above the Maeotian-Pontian contact. The Pontian deposits are overlain by a 3-m-thick member of dark gray, slightly brownish clays with gypsum, which was previously considered to represent the Pontian-Kimmerian transitional beds (Pevzner and Chikovani, 1978). Above the clays there are a thin (0.2 m) layer of dense, highly ferruginate marl that has been considered by Chumakov (2000) as a fossil soil and a 0.5-m-thick bed of dark gray clay overlain by a 3.5-m-thick limonite layer. Upsection, the overlying Kimmerian sediments are represented by a sequence of dark gray clays sandy in their upper part. The studied part of the sequence is 34-m thick. The lower 9 m of the Pontian deposits show normal polarity, while in the overlying 114-m-thick interval their polarity is reversed (Fig. 3).

Taking the units ages defined in the magnetochronological scale CK95 and increasing by 10% the fission-track date for the top of the lower Pontian, we can calculate now the Pontian time span and ages of its boundaries in the Zheleznyi Rog Cape section. Since the section is composed of uniform facies, we admit a constant sedimentation rate. The boundary between Pontian deposits with normal and reversed polarities is correlated with the basal limit of Chron 6 (C3Br) that is 7.43 Ma old according to the CK95 scale. The fission-track date of 7.0 Ma is selected for the upper boundary of the lower Pontian (for base of the *Congeria subrhomboidea* Beds). Hence, the lower Pontian deposits with reversed polarity, which are 64 m thick, accumulated during 0.39 m.y. If the sedimentation rate was constant, we may calculate that the 9-m-thick interval, the lower Pontian deposits have normal polarity, accumulated during 0.06 m.y., and the upper Pontian sediments, which are 50 m thick and show reversed polarity, were deposited during 0.30 m.y. The calculations suggest also that the Pontian base and top are 7.49 and

6.74 Ma old, respectively, in the Zheleznyi Rog Cape section. Thus, in this section, the Pontian base corresponds in age to 7.5 Ma, the upper Pontian base to 7.0–7.1 Ma, and the Pontian top is ca. 6.7 Ma old. According to mammalian fauna, the latter is estimated to be older than 6.57 Ma. It should be noted, however, that the fission-track date, being increased by 10%, implies a slightly older age of 7.7 Ma for the Pontian lower boundary. Accordingly, the Pontian time span in the Eastern Paratethys is 0.8 to 1.0 m.y. long.

Zubakov (1990) suggested that in the Zheleznyi Rog Cape section there is a hiatus in the range of Chron 5 (C3An) at the level of the *subrhomboidea* Beds. It is possible to calculate for such a case that 64 m of the lower Pontian sediments with reversed polarity accumulated during 0.86 m.y. (= duration of Chron 6), while 50 m of the upper Pontian deposits with reversed polarity were deposited (according to sedimentation rate) during 0.67 m.y. According to these calculations, the Pontian upper boundary should be 5.22 Ma old and would be falling into the lowermost Thvera Subchron. The last inference contradicts not only the paleomagnetic records, as the uppermost Pontian strata have reversed polarity, but also the data on the Kimmerian sediments between the Thvera Subchron and the Pontian top, which are 45 m thick in Well 15, and the nannofossil and mammal evidence.

After an early publication (Semenenko and Pevzner, 1979), Semenenko comprehensively studied the Zheleznyi Rog Cape section to search for nannofossils and to verify boundaries of the Kimmerian regional substages. In the lower part of the 3-m-thick member of brownish clays previously referred to the Pontian-Kimmerian transitional beds, there were found shells of *Paradacna abichi* (R. Hoern.). The overlying bed of dense ferruginate marl yielded the early Kimmerian *Paradacna deformis* Ebers. and *Dreissensia iniquivalvis* (Desh.). According to Zhabrev and Buryak (1958), clays sandwiched between the ferruginate marl and a higher limonite layer bear remains of *P. deformis* Ebers. and *Pteradacna edentula* (Desh.). Therefore, the Pontian-Kimmerian boundary is within this "transitional" member. Molluscan shells typical of the Kamyshburun beds were found in the 3.5-m-thick limonite layer. One meter below the ferruginate marl, there was encountered *Ceratolithus acutus* Gartner et Bukry (Lyul'eva, 1989). As noted above, this species was found in the Azov regional substage of Well 15 at the level of 45 m above the substage base and immediately below the Thvera Subchron. Consequently, deposits of the Azov Substage are represented in the Zheleznyi Rog Cape section by their middle part, have small thickness about two meters, and the Pontian and Kimmerian are separated here by a hiatus. The Kimmerian clays of normal polarity, which overlie the limonite bed, can be correlated with the Nunivak or Cochiti subchrons of the Gilbert Chron, because the top of the Azov Substage in Well 15 is recorded between the Sidufjall and Nunivak subchrons. Accordingly, there is a stratigraphic hiatus between the Azov and Kamyshburun beds of the Zheleznyi Rog Cape section (Fig. 3).

AGE OF THE MAEOTIAN-PONTIAN BOUNDARY IN THE EASTERN PARATETHYS

As is evident from data presented above, age estimates obtained by different methods for the Maeotian-Pontian boundary are well consistent between each other (table). According to nannofossils, the Maeotian lower boundary is older than 9.4 Ma and younger than 10.7 Ma, while the fission-track method showed that it is younger than 10.2 Ma. Paleomagnetic records imply age of 9.6 Ma for this boundary, since it is placed at the beginning of Chron C4Ar.2n (Pevzner and Vangengeim, 1993). It is the last date that we regard as the correct age estimate for the Maeotian base.

According to nannofossils, the Maeotian-Pontian boundary is older than 7.4 Ma and younger than 8.7 Ma, whereas its age is defined at 7.7 Ma by fission-track method and at 7.5 Ma by paleomagnetic data. When evaluating this boundary age, we prefer that of the paleomagnetic scale, because the fission-track date is obtained with a significant error. Mammalian fauna suggests that the lower-upper Pontian boundary age is close to 7.1 Ma, and fission-track method determines it at 7.0 Ma. From the standpoint of age determination accuracy, these dates are coincident. Unfortunately, the available data are inadequate for dating of the Pontian upper boundary and only restrict the upper limit of the stage. The limitation level is older than 5.6 Ma in terms of nannofossil chronology and older than 6 Ma according to mammal fauna. If the mammal fauna age is calibrated with reference to the paleomagnetic records at the Hatvan locality, the stage upper limit is older than 6.57 Ma, whereas interpretation of paleomagnetic characteristics of the Zheleznyi Rog Cape section suggests its position at 6.7 Ma. Since the lowermost part of the Azov regional substage is missing in the Zheleznyi Rog Cape section, it is likely that the uppermost Pontian deposits are also eroded there. Thus, the Pontian upper boundary is estimated to be 6.6–6.7 Ma old.

The estimated dates and age (6.2 Ma) of the upper boundary of the Shemakhan regional substage, the middle one in the Pontian of the Caspian basin, allow us to conclude that not only the upper Babadzhan, but also an upper half of the Shemakhan Substage of the Caspian basin corresponds to a part of the Kimmerian (Fig. 2).

The concrete inferences from the aforesaid are as follows. (1) The Pontian of the Eastern Paratethys has a "short" time span. (2) The stage corresponds to the uppermost Chron 7 coupled with a greater interval of Chron 6 of the magnetochronological scale. (3) It is impossible to correlate the Pontian with the lower part of the Gilbert Chron and upper part of Chron 5.

CORRELATION OF THE PONTIAN OF THE EASTERN PARATETHYS WITH THE MEDITERRANEAN STRATIGRAPHIC SCALE

Figure 2 shows correlation of mammal zones and regional stages of the Black Sea and Caspian basins, as we understood it based on materials discussed above, with the Mediterranean stages and nannofossil zones calibrated versus the magnetochronological scale after Berggren *et al.* (1995).

First, we should point to a close position of the Messinian base and the lower Pontian top near the time mark of 7–7.1 Ma. This level also corresponds to the beginning of the global C-isotope excursion. In the Ain El Beida section, Morocco, the latter event is recorded within the Chron 3Bn (Benson and Rakic-El Bied, 1996). It should be noted also that there was a sea-level drop by 150-200 m in the Late Miocene, between the base of Chron 3Bn and mid-Chron 3r (Harland *et al.*, 1985). The time span of this event almost coincides with the range of the Messinian. In the scale CK95, its lower and upper limits correspond in age to 7.1 and ca. 5.5 Ma, respectively. The lower one coincides with the Messinian base and the lower Pontian top in the Eastern Paratethys. It is reasonable to suggest therefore that isolated Black Sea and Caspian basins appeared instead of a single Pontian basin in response to that global sea-level drop, which also triggered the runoff of Pontian waters in the Mediterranean region. The last event is a good explanation for appearance of the Caspian-type mollusks and ostracodes in the Messinian sediments.

Rather widespread brackish-water sediments bearing the Caspian-type fauna are well known in the Mediterranean areas under names Lago-Mare, Colombacci Formation, *Congeria*, *Melanopsis*, and *Loxoconcha djaffarovi* beds. Many researchers (Popov and Nevesskaya, 2000, and references therein) who studied the mollusks and ostracodes from these deposits correlate the latter, partially or completely, with the Pontian of the Eastern Paratethys.

Even if the Lago-Mare sediments correspond indeed to a part of the Pontian in the Eastern Paratethys, it would be incorrect to evaluate the Pontian position in the magnetochronological scale based on the age of Lago-Mare subdivision, as it has been done in some works (Popov and Nevesskaya, 2000; Popov *et al.*, 2001). In our opinion, it is more correct to define the age of Lago-Mare sediments from the Pontian position in the magnetochronological scale, since the available age determinations for these sediments are not convincing. Those few dates (Krijgsman *et al.*, 1999), which we found, indicate an older age of Lago-Mare strata (5.50–5.33 Ma) than it used to be accepted.

For instance, the autochthonous brackish-water ostracodes of the Caspian-type were encountered in the Cuevas de Almanzora section of the Vera basin, Spain, in the Messinian deposits of the *D. quinqueramus* Zone NN11 (Barragan *et al.*, 1990; Benson and Rakic-El Bied, 1996). Their host deposits underlie sediments bearing an early evolutionary form *Globorotalia margaritae* Bolli et Bermudez whose FAD is recorded in the middle of Chron C3An2n at the level of about 6.35 Ma (Benson and Rakic-El Bied, 1996). Consequently, the deposits with brackish-water ostracodes are even older.

In DSDP Hole 372 drilled on the Menorca Rise, the Messinian sediments, which span the interval of Zone NN11 or of the upper *G. plesiotumida* Zone N17 (cores 4-9), unconformably overlie the Serravalian deposits and are overlain with a hiatus by the Pliocene sediments of the *Globorotalia puncticulata* Zone MPL3. *Cyprideis pannonica* (Mehes) characteristic of the Lago-Mare sediments was found here in the Core 4-2, 90–92 cm (*Initial Reports...*, 1978). If the Lago-Mare strata are within zones NN11 and N17, then they are older than 5.6 Ma (LAD of *Discoaster quinqueramus* Gartner) or than 6.35 Ma (FAD of *Globorotalia margaritae* Bolli et Bermudez).

The third section important for the age determination of the Lago-Mare sediments is located in northern Apennines. In the Monticino gypsum quarry, Brisighella, Ravenna, the 2.3-m-thick Colombacci Formation composed of sediments with reversed polarity overlies, with a stratigraphic and angular unconformity, the gypsum beds of the Gessoso-Solfifera Formation (Marabini and Vai, 1988; Vigliotti, 1988). These sediments fill in cracks in the underlying gypsum and yield the mammal fauna of Zone MN13 (Rook, 1992). Since the upper boundary of Zone MN13 is 6 Ma old, the bone beds of the Colombacci Formation are older than this date.

The dates cited above do not contradict the commonly accepted correlation of the Lago-Mare sediments and their Mediterranean analogs with the Pontian of the Eastern Paratethys. These sediment correspond only to the upper Pontian or to a part of the latter, and are not younger than 6.6–6.7 Ma (top of the Pontian in the Eastern Paratethys). When the Lago-Mare sediments are overlain by deposits of the Zanclean transgression, we should admit a stratigraphic hiatus between them that is over one million year long.

CONCLUSIONS

The discussed materials lead to the following conclusions.

(1) The Pontian of the Eastern Paratethys corresponds to the uppermost Chron 7 coupled with a greater portion of Chron 6 and ranges in age from 7.5 to 6.7–6.6 Ma.

(2) The lower Pontian is correlative with the upper-part of the Tortonian, where sediments show reversed polarity.

(3) The beginning of the late Pontian coincides with the beginning of the Messinian, and the late Pontian

corresponds to the first third of the Messinian, if the latter ranges in age from 7.1 to 5.33 Ma.

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The Pliocene-Pleistocene Golovnin Formation of Kunashir Island (the Kuriles): Stratigraphy and Formation Conditions

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Abstract—Lithology and facies of shelf deposits of the Golovnin Formation have been studied in the Kunashir Island. The deposits may be regarded as an intermediate link between correlative sediments of marginal seas and the Pacific Ocean. Based on diatoms, spores, pollen, and paleomagnetic analysis, the Golovnin Formation is subdivided into three subformations representing different transgression-regression cycles of the terminal Pliocene-Pleistocene. Significant hiatuses have been established in formation of the volcanogenic-sedimentary sequence. The *Pyxidicula zabelinae* and *Thalassiosira antiqua* zones of the upper Pliocene are distinguished in deposits of the lower subformation. The middle subformation that yields diatom assemblages of the *Actinocyclus oculatus* and *Nitzschia fossilis* zones formed in the early Pleistocene, whereas deposits of the upper subformation belong to the diatom *Proboscia barboi* Zone of the middle Pleistocene. Terrestrial deposits are united into the Belozersk Beds of the second late Pleistocene interstadial. Two, the Kunashir I (2.12–2.0 Ma) and Kunashir II (1.8–1.6 Ma) transgression phases of the late Pleistocene are distinguished. The early Golovnin transgression occurred in the early Pleistocene. The late Golovnin transgression (0.43–0.28 Ma) with two phases corresponding to oxygen-isotope stages 11 and 9 complies with the middle Pleistocene. Tephrostratigraphic markers are established for both the diatom zones and the Reunion and Olduvai subchrons. Paleogeographic events in southern Kuriles, Japan, Kamchatka, Chukchi Peninsula, and Alaska are correlated between each other.

Key words: Late Cenozoic, Kuriles, stratigraphy, paleogeography.

INTRODUCTION

Characteristic of the southern Kunashir Island are widespread Pliocene and Pleistocene volcanogenic-sedimentary sequences representing stratigraphic range of the Golovnin Formation, the stratotype of which was distinguished on the southeastern island coast (Piskunov, 1963). The formation is composed of tuff-silt, tuff-diatomite, and sands alternating with ash interlayers and tephroid members. It is overlain by eolian sandy loam, alluvial and lacustrine-alluvial pebble beds and sands of the upper Pleistocene and Holocene. High sedimentation rates determined by intense volcanism resulted in accumulation of thick sedimentary strata, which recorded volcanic stages, abrupt paleoclimatic changes, and sea-level fluctuations. Deposits of the Golovnin Formation represent a key object important for stratigraphic correlation of deposits of marginal seas and NW Pacific, and of stratigraphic subdivisions established in lower and higher latitudes.

Marine sequences of the Golovnin Formation of southern Kunashir (up to 350 m thick), which rest on rocks of the Alekhino Formation being overlain by continental facies (up to 70 m thick), are traceable in the Golovnin Cliff (up to 120 m high) from the Cape Puzanov to Belozerskaya River mouth (Fig. 1). Based on the molluscan fauna, the formation was referred to the late Pliocene or late Pliocene-Pleistocene (Piskunov, 1963; Zhidkova *et al.*, 1974). Later on (Pushkar' *et al.*, 1998a, b), the formation was divided into three subformations, and continental facies were attributed to the independent Belozersk Beds.

The purpose of the work is stratigraphic differentiation of deposits of the Golovnin Formation in its stratotype. Other objectives of the study are determining the depositional environments and distinguishing the paleogeographic events, which could be indicative of global environmental changes during the Late Cenozoic.

MATERIALS AND INVESTIGATION METHODS

Deposits of the Golovnin Formation were studied in its stratotype, in a series of coastal exposures of the Izmena Bay (from the Khlebnikov Creek mouth to the Golovnin River mouth) and Sea of Okhotsk (from Ivanovskii Cape to the Temnaya River mouth), and in the section of 40-m-high terrace-like structure of the South Kurile Isthmus (Fig. 1).

Lithology of the Golovnin Formation deposits, the tephra beds included, has been studied in detail necessary for distinguishing lithostratigraphic units. The main index fossils in deposits are diatoms whose well-preserved valves are abundant in sediments. By age

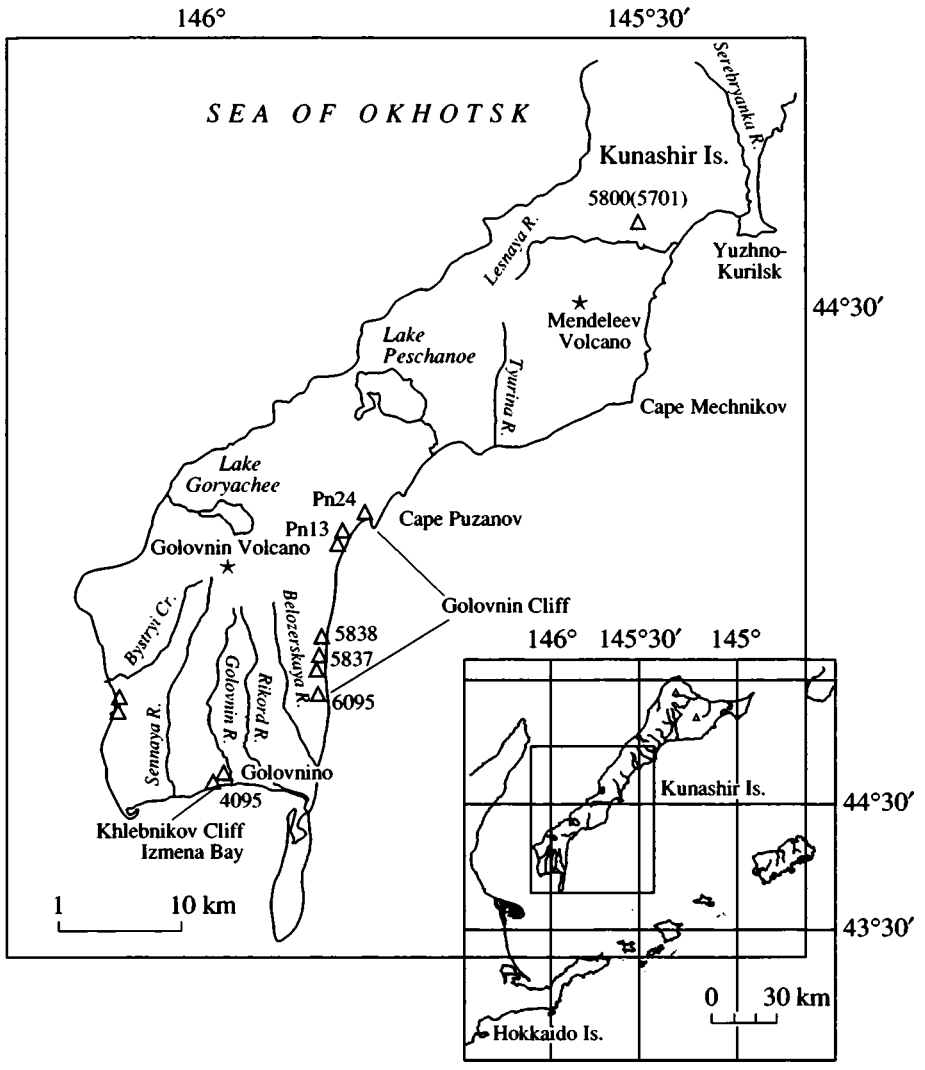


Fig. 1. Localities of the Golovnin Formation sections studied in the Kunashir Island.

interpretation of diatom assemblages, the modified North Pacific diatom zonation (Fig. 2) was used. The zonation is based on the classical scheme (Koizumi and Tanimura, 1985) and data on synchronism and diachronism of diatom evolutionary levels in different regions of Central and North Pacific (Barron, 1992a; Barron and Gladenkov, 1995; Pushkar *et al.*, 1999). Spores and pollen have also been studied, and paleomagnetic analysis has been carried out in addition. To identify paleomagnetic events, models of changes in the magnetic field intensity of the Earth for the past 3 Ma were used (Berggren *et al.*, 1995; Raymo, 1992; Cande and Kent, 1995).

The Pliocene-Pleistocene boundary is positioned at the upper level of the Olduvai Subchron C2n (1.69 Ma), or slightly above it. The boundary between the early and middle Pleistocene is considered at the level separating the Matuyama and Brunhes chrons, C1r1r/C1n (0.79 Ma). The boundary between the middle and late Pleistocene coincides with the beginning of phase "e" 5 of the oxygen-isotope stage (OIS) dated at 130 ka (Williams *et al.*, 1988; Berggren *et al.*, 1995; Zijderveld *et al.*, 1991).

LITHOLOGY

The stratotype of the Golovnin Formation represents a volcanogenic-sedimentary sequence making up the dome structure of Golovnin Volcano and having a gentle dip in the frontal part and monoclinal dipping on flanks. The structure is complicated by dislocations and seismic faults (Fig. 3). Lower horizons of the Golovnin Formation are most completely represented in the Khlebnikov Cliff section (Fig. 4).

In the Golovnin Formation, the following lithological members characterized by lateral facies changes are distinguished:

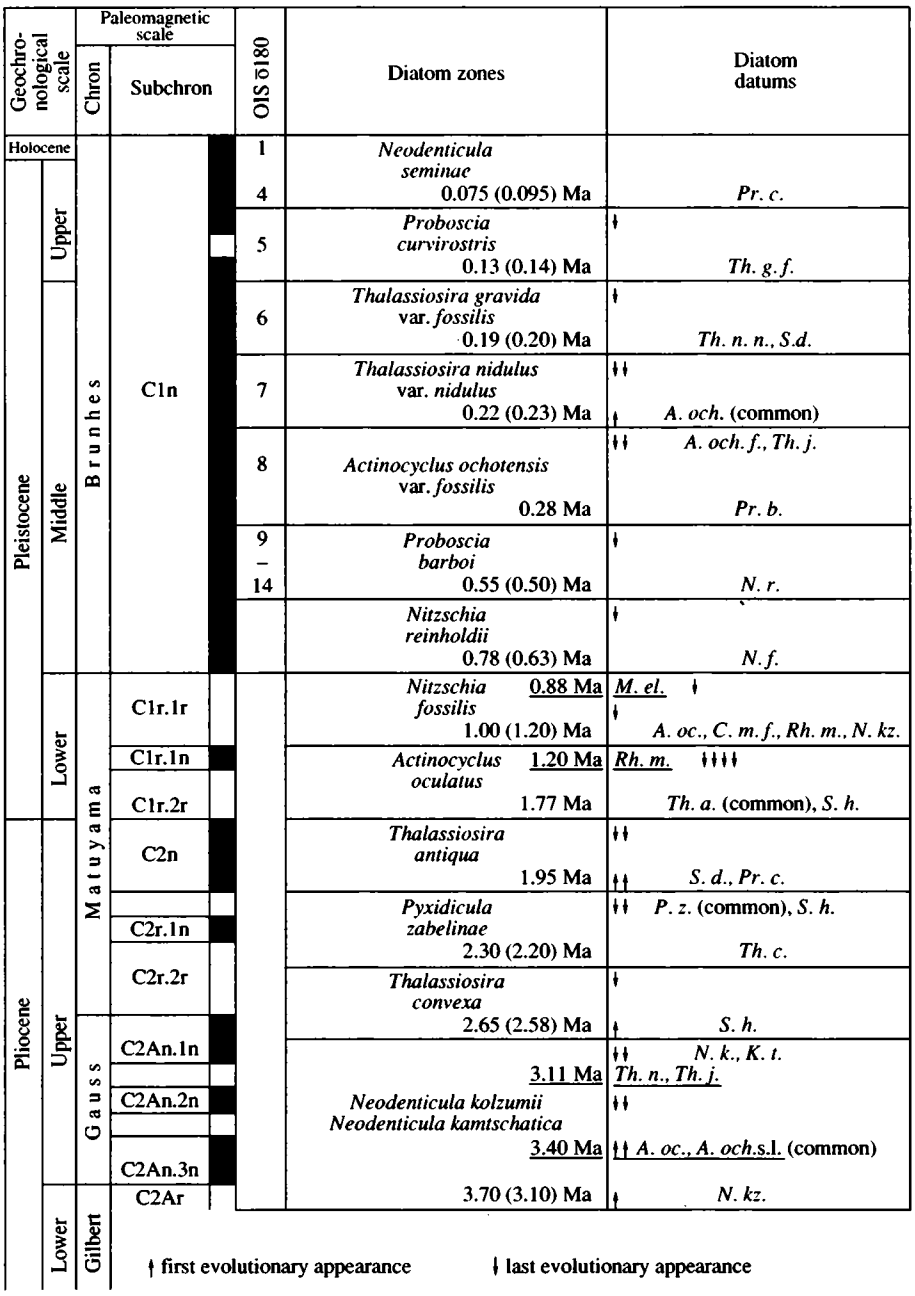


Fig. 2. Diatom zonation and datums established in NW Pacific (Pushkar *et al.*, 1999): (N.kz.) *Neodenticula koizumi* Akiba et Yanagisawa; (A.oc.) *Actinocyclus oculatus* Jousé; (A.och.) *Actinocyclus ochotensis* Jousé; (A. och.f.) *Actinocyclus ochotensis* Jousé var. *fossilis* Pushkar; (Th.n.) *Thalassiosira navita* Sheshukova; (Th.j.) *Thalassiosira jacksoni* Koizumi et Barron; (N.k.) *Neodenticula kamtschatica* (Zabelina) Akiba et Yanagisawa; (K.t.) *Koizumia tatsunokuchiensis* Yanagisawa; (S.h.) *Stephanopyxis horridus* Koizumi; (Th.c.) *Thalassiosira convexa* Mukhina; (P.z.) *Pyxidicula zabelinae* (Jousé) Jordan et Priddle; (Th.a.) *Thalassiosira antiqua* (Grun.) Cl.; (Rh.m.) *Rhizosolenia matuyamai* Burckle, C.m.f.) *Coscinodiscus marginatus* Ehr. var. *fossilis* Jousé; (N.f.) *Nitzschia fossilis* (Frenguelli) Kanaya; (N.r.) *Nitzschia reinholdii* Kanaya; (Pr.b.) *Proboscia barboi* (Temp. et Brun.) Jordan et Priddle; (Th.j.) *Thalassiosira jouseae* Akiba; (Th.n.n.) *Thalassiosira nidulus* var. *nidulus* (Temp. et Brun.) Jousé; (Th.g.f.) *Thalassiosira gravida* Cl. var. *fossilis* Jousé.

Member 1, up to 26 m thick, is represented in the stratotype by gray, dense, thin-bedded siltstone. Beds dip eastward (dip azimuth 80°) at the angle of 5°–10°. In the southern part of the Golovnin and Khlebnikov cliffs, deposits of the lower member are represented by alternating pumiceous coarse sand and gravely deposits of underwater pyroclastic pumice flows characterizing two intense phases of the Golovnin Volcano activity. In the section of the Khlebnikov Cliff, two tephrostratigraphic markers were distinguished in the lower sub-

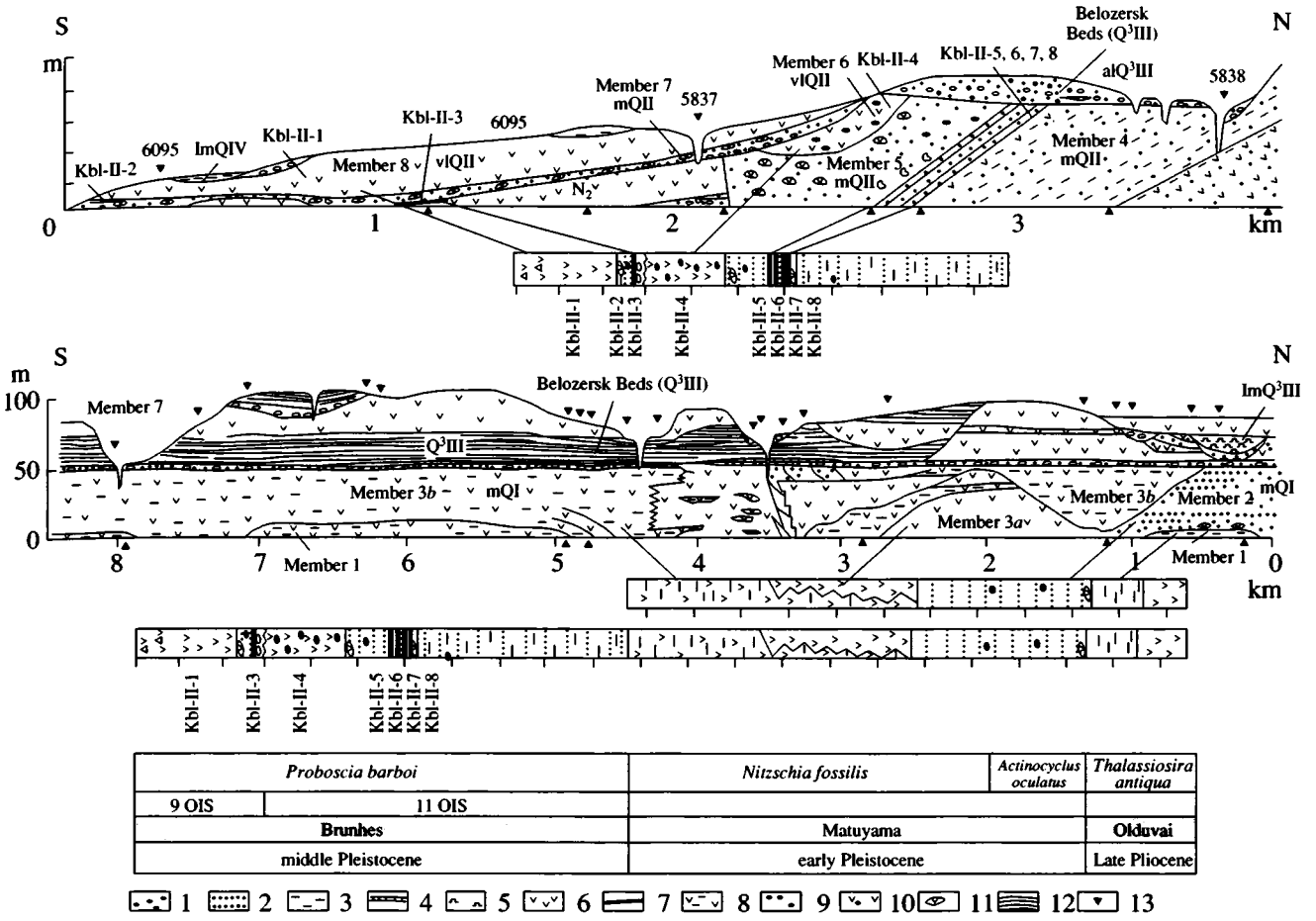


Fig. 3. Geological structure of the Golovnin Cliff: (1) pebble; (2) sand; (3) siltstone; (4) peat; (5) diatomite; (6) tuff; (7) volcanic ash; (8) tuff-siltstone; (9) pumice; (10) pumice tuff; (11) volcanic bomb; (12) terrestrial complex (Belozersk Beds); (13) studied sections.

formation: the well-traceable regional ash bed Kh1-3, the source of which was volcanism in central Hokkaido, and well-rounded pumice occurring at the base of the section (Razzhigaeva *et al.*, 2002).

Member 2, 37 m thick, is exposed in the northern Cape Puzanov site of the stratotype area. It is composed of yellowish, thin-bedded, inequigranular pumiceous sand with large volcanic bombs and underwater slump structures, which grades upward into ferruginate, medium-grained, well-graded sand with interlayers of pumice gravel and small pebbles. The upper contact is not exposed.

Member 3, up to 60 m thick, in the upper part of the stratotype is composed of light gray, thin-bedded tuff-siltstone grading into facies of dacite pumice sands, which represent deposits of sub-aqueous pyroclastic flows of Golovnin Volcano filling in depressions of old relief. The deposits exhibit marks of intense sub-aqueous slumping and seismic faulting. They accumulated along edges of the underwater volcano under conditions of high seismicity. In the upper part, siltstone grade into pumice tephroids, the gray-green inequigranular sand with gravel and well-rounded pumice otherwise defined as obliquely laminated sand of 700000 ± 210000 years old (Bulgakov, 1994).

Member 4, 45 m thick, is exposed in the southern part of the cliff and shows monoclinal dipping. It is represented by alternating tephroid siltstone and inequigranular sand with well-rounded pumice gravel and pebble, and with lenses of tuff-siltstone and tuff-breccia at the top. Around the edges of Golovnin Volcano, deposits accumulated in the course of volcano tephra reworking. In the South Kurile Isthmus, coeval deposits enclose regional beds of volcanic ash (K_{yk}-II-5 and K_{yk}-II-4), the provenance of which was in Hokkaido (Razzhigaeva *et al.*, 2002).

Member 5, about 25 m thick, resting conformably on Member 4 is composed of greenish yellow, slightly lithified sandstone with thin lenticular lamination in the lower part emphasized by black sand laminae, and with horizontal ripple lamination in the upper part. The found shells of marine mollusks represent taxa *Elliptica* cf. *alaskensis derbeki* Scarlato, *Tridonta* ex gr. *borealis borealis* (Schumacher), *Felaniella* cf. *usta*

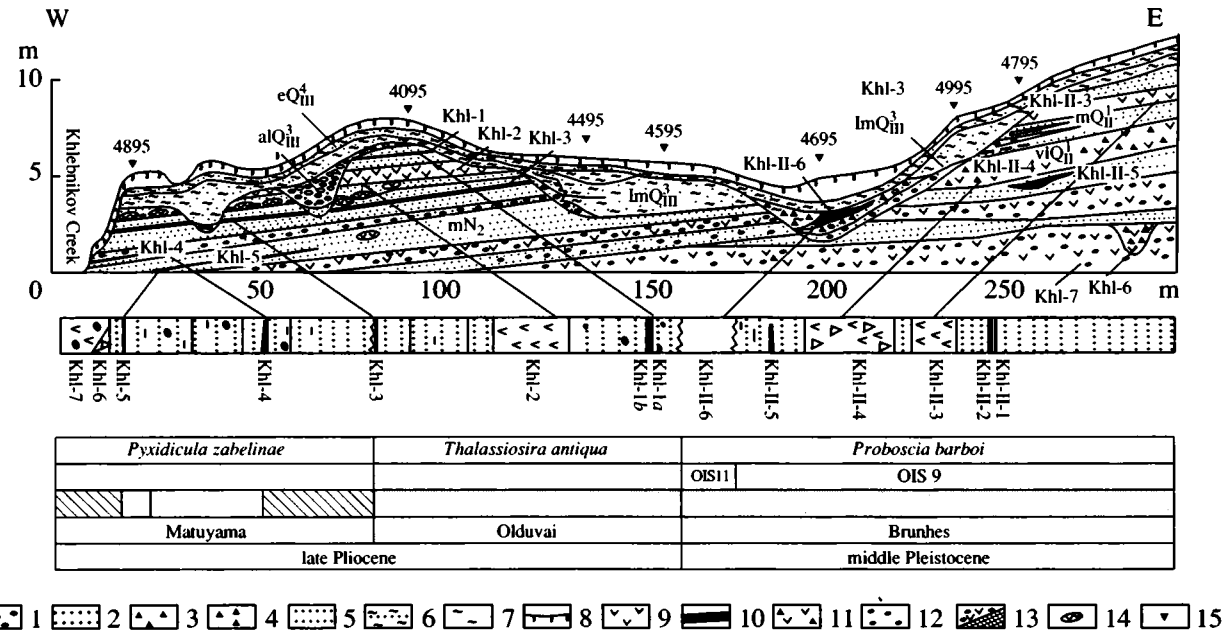


Fig. 4. Geological structure of the Khlebnikov Cliff, Izmena Bay: (1) pebble; (2) gravel; (3) rock debris; (4) grus; (5) sand; (6) sandy loam; (7) loam; (8) soil; (9) tuff; (10) volcanic ash; (11) tuff breccia; (12) pumice; (13) pumice tephroid; (14) volcanic bombs; (15) studied sections (indices Khl-3 etc. indicate tephra interlayers).

(Gould), *Yagudinella* ex gr. *notabilis* (Sowerby) (identified by V.D. Khudik). The lower part of the member encloses four rhyolite ash interlayers, the upper one K_{bl}-II-8 with a high K_2O content being regional and probably related to eruptions of Hokkaido volcanoes (Razzhigaeva *et al.*, 2002). The upper part of the member is rich in pumice, volcanic glass and coarse resurgent material, which mark the beginning of a new cycle of volcanism, high seismicity, and large blocks displacement on the continental slope (Fig. 2).

Member 6, about 16 m thick, corresponds to tephroids composed of well-rounded white dacite pumice. Its deposits recorded termination of next volcanic cycle.

Member 7, up to 6 m thick, is represented by beds of beach pebbles and boulders, which grade into yellowish gray tephroid sands. The member overlies, with erosion and stratigraphic hiatus, the underlying deposits of members 2-6. In the southern part of the Golovnin Cliff, the sands enclose a distinct and well-traceable bomb horizon in their upper part, which is a lithological marker of the member.

Member 8, 20 m thick, is composed of dacite to rhyolite pumice tuff and tuff-breccia K_{bl}-II-1 conformably resting on sands and marking a new phase of the Golovnin Volcano activity during a sea-level drop. The deposits, along with those of Member 7, seem to fill in the U-shaped valley in the southeastern part of the island.

The terrestrial deposits of the studied section, which are assigned to Belozersk Beds, represent a complicated sedimentary cover composed of pyroclastic flows and sheets, volcanic aprons, and dry river fans, which are intercalated with interlayers of volcanic ash, peat bogs, alluvial and lacustrine facies (Razzhigaeva *et al.*, 2000). A series of radiocarbon dates obtained for deposits is within the range of 48–31 ka. The section top is composed of yellow-brown massive to porous sandy loam formed probably in the terminal Pleistocene-early Holocene.

STRATIGRAPHY

Diatom assemblages are studied from deposits of all the members and Belozersk Beds. No traces of redeposition have been revealed, but old *Aulacoseira* forms found in assemblages indicate that marine deposits of the island associate with Miocene or Pliocene terrestrial deposits. Based on the mineral and chemical analyses, and on stratigraphic range of diatoms, deposits of the Golovnin Formation can be subdivided into three subformations.

Lower Subformation (Member 1) was formed during the first volcanic stage. Its deposits are represented most completely in the section of the Khlebnikov Cliff (Section 4095; Fig. 5), where two diatom assemblages are distinguished.

Assemblage 1 (6.9–4.2 m) includes 135 species of 50 genera. The abundance of valves is as high as 2 to 3 millions per one gram of sediment. The dominant taxa are *Thalassiosira gravida* var. *fossilis, Th. gravida* var. *gravida, Th. eccentrica, Actinocyclus curvatulus* + *A. divisus*, and spores *Chaetoceros* spp. Abundant are sublittoral diatoms *Paralia sulcata, Hyalodiscus obso-*

letus and species of the genus *Diploneis*: *D. smithii, D. interrupta, D. subcincta.* In terms of age relations, the assemblage corresponds to the *Pyxidicula zabelinae* Zone (Fig. 5). It is confined to sediments of the Matuyama Chron of reversed polarity. Sediments of the normal polarity interval at the base of the diatom zone likely correspond to one of the excursions of the Cronozone C2r1n of the Reunion Subchron, most likely to the upper one 2.08 Ma old.

Assemblage II (4.2–1.4 m) is similar in taxonomic composition to Assemblage 1 but differs greatly in the abundance of valves and number of age groups (Fig. 5). The abundance of diatoms does not exceed 0.5 million valves per one gram of sediments. Species dominating in the assemblage are *Thalassiosira gravida* var. *gravida* and *Th. gravida* var. *fossilis.* The group of Pliocene diatoms is represented by single valves, whereas the total abundance of Pleistocene species proper increases. Representatives of the *Thalassiosira* genus are common among the latter. The assemblage corresponds in age to the *Thalassiosira antiqua* Zone. A high diversity, but low occurrence frequency of diatom species, reflects most likely the initial periods of the assemblage formation. The *Thalassiosira antiqua* Zone of normal polarity is correlative with the Olduvai Subchron C_2n. This is consistent with found silicoflagellate species *Dictyocha subarctios* Ling, the appearance of which in marine sediments of the North Pacific is an important marker of the Olduvai Subchronozone upper boundary (Ling, 1976).

Similar diatom assemblages are distinguished in Section 4895 (3.75–2.15 m) and in dense siltstone beds with normal polarity occurring in the lower part of the Golovnin Formation at the Cape Puzanov site (Fig. 3). The *Thalassiosira antiqua* Zone is an important marker in the diatom evolution, because many Neogene species disappear here and Pleistocene species of the genus *Thalassiosira* become abundant.

Several ash interlayers of Khl series were distinguished in deposits (Razzhigaeva *et al.*, 2001). Source volcanism for the K_{hl}-3 ash bed was in Hokkaido, where ash interlayers with similar chemical and physical characteristics mark the Olduvai Subchronozone inside the Ikeda and Uranosawa formations (Kitagawa *et al.*, 1987). A bed of well rounded pumice below the K_{hl}-5 ash layer is correlative with the late Pliocene Chiyoda pumice flow in Hokkaido (Pushkar' *et al.*, 1986b), which marks the Reunion Subchronozone (Kitagawa *et al.*, 1982).

Middle Subformation (Members 2 and 3). Diatoms dominating in the lower part of Member 2 are *Thalassiosira gravida* var. *fossilis, Stephanopyxis horridus* and *Coscinodiscus marginatus* var. *fossilis*, which coexist with some forms transitional to the recent *Coscinodiscus marginatus.* Valves of *Actinocyclus oculatus* are rare (one valve in two-three rows at magnification ×600), but they imply that these deposits can be conventionally attributed to the synonymous zone. Diatoms in sands of the upper part of the member are also rare, but their assemblage is different, represented by *Thalassiosira gravida* var. *gravida, Th. eccentrica, Th. sancettae, Th. antarctica, Th. latimarginata, Thalassiosira nidulus* var. *nidulus, Th. jouseae, Porosira glacialis*, and *Odontella aurita* associated with species of the genus *Nitzschia, N. fossilis* included. Accordingly, the assemblage is characteristic of the *Nitzschia fossilis* Zone. In the upper part of the member, dominant neritic forms are replaced by shallow-water benthic and epiphytic flora represented by species of genera *Diploneis, Cocconeis, Trachyneis. Odontella aurita* and *Nitzschia grunowii* dominating amidst neritic diatoms are indicative of cooling. In the Cape Puzanov, the Belozersk Beds situated above the level of 45 m enclose a thick (up to 10 m) dolomite lens with freshwater *Stephanodiscus-Aulacoseira* flora (middle ?–late Pleistocene), and also siltstone and sand interlayers with *Fragilaria incognita, F. nana* and *Nitzschia nana* flora.

The assemblage from the *Nitzschia fossilis* Zone of reversed polarity Member 3 (an upper part of the Matuyama Chron) that is exposed over the distance of 4.3–8.5 km along the Golovnin Cliff is similar in composition to that of Member 2 from the Cape Puzanov site though different in valve abundance (4–5 millions per a gram of rock). Especially abundant are *Stephanopyxis dimorpha* (11.3%) and species of the genus *Thalassiosira*: *Thalassiosira gravida* var. *fossilis, Th. antarctica, Th. nidulus* var. *nidulus* + *Th. jouseae* and *Th. sancettae.* Representatives of *Neodenticula seminae* and diversified forms of *Proboscia barboi* and *P. curvirostris* are common. In upper tuff-sandstones of the member, the role of sublittoral species increases (up to 17.8%), and *Paralia sulcata* becomes dominant (18.4%). The index species abundance in tuff-siltstone is discrete, and only individual representatives (3–5 per a preparation) were encountered in tuff-sandstone.

Upper Subformation. Deposits of the upper subformation of the Golovnin Formation (Members 4–8) were studied in its stratotype (Fig. 3), in the section of the Khlebnikov Cliff (Fig. 4), and in the 40-m-high terrace-like structure of the South Kurile Isthmus (Pushkar' *et al.*, 1998a, b). Based on paleomagnetic data, the deposits are referred to the Brunhes Chron.

In the Golovnin Formation stratotypes, diatoms of the *Proboscia barboi* Zone were distinguished in deposits of members 4 (Section 5838) and 7 (Sections 5837 and 6095). Diatoms of Member 5 were studied from samples collected from the lower cliff part (Fig. 3). Diatoms of the *Proboscia barboi* Zone are most characteristic of Section 6095.

Thalassiosira gravida var. *gravida* (14%) and *Bacterosira fragilis* (10%) are dominating diatom forms in Section 5838, in its lower part (9.10–7.8 m). Diatoms are rare in the middle part of the section (7.8–5.0 m), whereas stenohaline plankton prevails (up to 65%) in the upper part of the section (5.0–2.0 m). The assemblage reflects a climax state of the paleocommu-

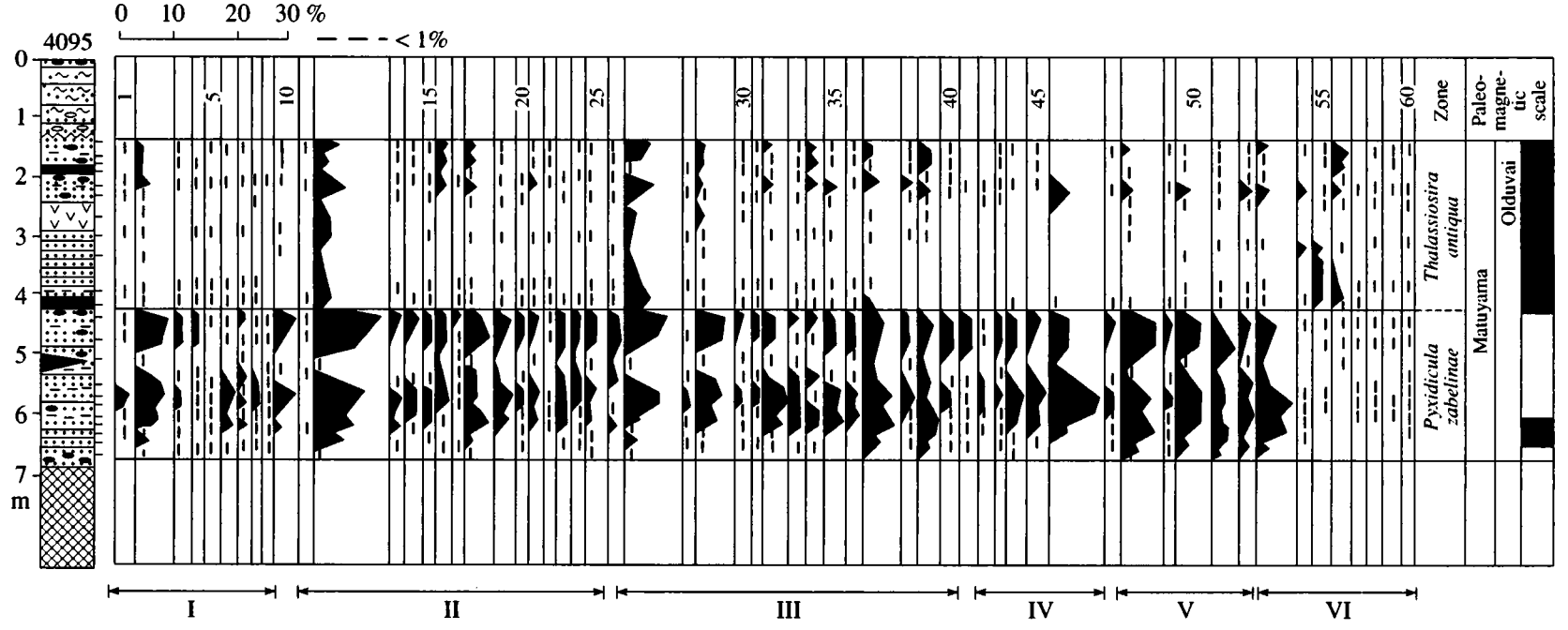


Fig. 5. Distribution of diatoms in Section 4095, Khlebnikov Cliff (abundance rate of 1% means 3 valves; symbols as in Fig. 4).

Group I (species died out in the late Pliocene): 1, *Stephanopyxis horridus* Koizumi; 2, *Pyxidicula zabelinae* (Sheshukova) Streln. et Nikolaev; 3, *Thalassiosira antiqua* (Grun.) Cl.; 4, *Th. tertiaria* Sheshukova; 5, *Th. manifesta* Sheshukova; 6, *Coscinodiscus pustulatus* Mann; 7, *Xantiopyxis diaphana* Forti; 8, *Synedra jouseana* Shushukova f. *linearis*; 9, *Cymatosira debyi* Temp. et Brun.

Group II (taxa died out in the Pleistocene): 10, *Stephanopyxis dimorpha* Schrader; 11, *S. californica* Schrader; 12, *Thalassiosira gravida* Cl. var. *fossilis* Jousé; 13, *Th. nidulus* Jousé var. *nidulus*; 14, *Th. jouseae* Akiba; 15, *Th. haynaldiella* Jousé + *Porosira (Thalassiosira) punctata* (Jousé) Makar.; 16, *Th. sheshukovae* Makar.; 17, *Paralia polaris* (Grun) Gleser; 18, *Coscinodiscus marginatus* Ehr. var. *fossilis* Jousé; 19, *Actinocyclus ochotensis* Jousé var. *fossilis* Pushkar; 20, *A. oculatus* Jousé; 21, *Proboscia (Rhizosolenia) barboi* (Brun.) Jordan; 22, *P. curvirostris* (Jousé) Jordan; 23, *Neodenticula koizumii* (Simonsen) Akiba et Yanagisawa; 24, *Nitzschia extincta* Kozyr. et Shesh.; 25, *Trochosira spinosa* Kitt.

Group III (transit oceanic to neritic Arctic- and north-boreal forms): 26, *Stephanopyxis turris* (Grev. et Arn.) Ralfs + var. *cylindrus* Grun.; 27, *Thalassiosira gravida* Cl. var. *gravida*; 28, *Th. hyalina* (Grun.) Gran; 29, *Th. eccentrica* (Ehr.) Cl.; 30, *Th. kryophila* (Grun.) Jorg.; 31, *Th. nordenskioeldii* Cl.; 32, *Porosira glacialis* (Grun.) Jorg.; 33, *Melosira arctica* (Ehr.) Dickie in Pritchard; 34, *Bacterosira fragilis* Gran; 35, *Coscinodiscus marginatus* Ehr. var. *marginatus*; 36, *Actinicyclus ochotensis* Jousé var. *ochotensis* Jousé var. *ochotensis*; 37, *Chaetoceras* sp. sp. (resting spores); 38, *Rhizosolenia setigera* Bright. var. *arctica* I. Kiss ?; 39, *Rhizosolenia hebetata* (Bail.) Gran f. *hiemalis* Gran; 40, *Pseudopyxila americana* (Ehr.) Forti; 41, *Nitzschia cylindra* (Grun.) Hasle.

Group IV (transit oceanic to neritic south-boreal and subtropical forms): 42, *Thalassiosira oestrupii* (Ostf.) Hasle; 43, *Th. decipiens* (Grun.) Jorg.; 44, *Coscinodiscus asteromphalus* Ehr. var. *asteromphalus*; 45, *C. radiatus* Ehr. + *C. perforatus* Ehr.; 46, *Actinocyclus curvatulus* Janish var. *curvatulus* + *A. divisus* (Grun.) Hust.; 47, *Stellarima microtrias* (Ehr.) Hasle et Sims.;

Group V (transit sublittoral north-boreal forms): 48, *Paralia sulcata* (Ehr.) Cl. var. *sulcata*; 49, *Radialiplicata sol* (Ehr.) Gleser; 50, *Hyalodiscus obsoletus* Shesh.; 51, *Odontella aurita* (Lyngb.) Ag.; 52, *Trachyneis aspera* Cl. var. *aspera*; 53, *Diploneis smithii* (Breb.) Cl. var. *Smithii* + *D. interrupta* (Kuetz.) Cl. var. *interrupta* + *D. subcincta* (A. S.) Cl.

Group VI (forms appeared in the Pleistocene): 54, *Thalassiosira pacifica* Gran et Angst.; 55, *Th. latimarginata* Makar.; 56, *Th. antarctica* Comber; 57, *Th. plicatoides* (Simonsen) Akiba; 58, *Th. sancettae* Akiba; 59, *Asteromphalus robustus* Castr.; 60, *Neodenticula seminae* (Simonsen) Akiba et Yanagisawa.

nity. In Section 5837, extinct diatoms are scarce, and dominating forms are *Paralia sulcata, Trachyneis aspera, Nitzschia cylindra* + *N. grunowii, Odontella aurita, Diploneis interrupta, D. subcincta.* The assemblage is indicative of more shallow-water setting as compared to that of the coeval assemblage from Section 6095. Only *Proboscia barboi* makes up 2%. Dominating in the assemblage (9.5–6.6 m) is *Paralia sulcata* var. *biseriata* (86.8%) that probably became extinct in the middle to initial late Pleistocene time.

In Section 4695 of the Khlebnikov Cliff, the diatom assemblage of *Proboscia barboi* Zone consists of 122 taxons (Pushkar' *et al.*, 1998b) and is similar to the assemblage from Section 5838. In Sections 4795 (6.0–5.5 m) and 4995 (3.3–4.6 m) of the same cliff, there are assemblages with scarce zonal species *Proboscia barboi*, and abundant *Paralia sulcata* var. *biseriata* (up to 26.8%), which are comparable with the assemblage of Section 5837.

In the South Kurile Isthmus, there is a member of tuffaceous diatomite ($SiO_{2amorph}$ = 74.3%) at the base of Section 5800 (5701) (9.8–13 m). The middle part of the section (5.0–9.8 m) is composed here of tephroid siltstone ($SiO_{2amorph}$ = 18.96%) and the upper one (1.0–5.0 m) of alternating siltstone and sand with pumice, which enclose interlayers K_{yk}-II-3, K_{yk}-II-2, and K_{yk}-II-1 of dacitic ash.

The diatom assemblage includes species, which died out in the Pleistocene (up to 6.7%): *Thalassiosira gravida* var. *fossilis, Th. nidulus, Th. jouseae, Actinocyclus ochotensis* var. *fossilis, Proboscia curvirostris, P. barboi*, and *Stephanopyxis dimorpha.* In general, forms dominating in the assemblage are planktonic diatoms *Actinocyclus curvatulus* var. *curvatulus, A. divisus, Th. eccentrica*, and *Chaetoceras* spp. (resting spores), as well as sublittoral benthic and shallow-water planktonic diatoms *Delphineis kippae, Hyalodiscus obsoletus*, and *Paralia sulcata* var. *sulcata.* Upward in the section, there is recorded change of prevailing planktonic and benthic diatoms that is related to transgression phases. Found in the same deposits are silicoflagellates *Paradictyocha polyactis* (Ehr.) Freng. f. *competa* Freng., *Distephanus speculum* (Ehr.) Haeckel var. *speculum, D. speculum* var. *septenarius* (Ehr.) Jorg., *D. octonarius* (Ehr.) Defl. var. *polyactis* (Jorg.) Gleser, and *Dictyocha fibula* (Ehr.) var. *fibula* (7% in total). *Distephanus octonarius* var. *polyactis* found in the assemblage is an additional marker of the Mindel-Riss upper boundary (Jousé, 1969).

Deposits of the Golovnin Formation are everywhere overlain with stratigraphic hiatus by tuffaceous and lacustrine-alluvial deposits assigned by us to the Belozersk Beds. Recorded in the beds is alternation of cold-water benthic *Pinnularia borealis* var. *borealis, P. lata, P. episcopalis, P. streptoraphe* et vars, *P. viridis* et var. *intermedia, Eunotia bidens*, and planktonic warmer-water *Aulacoseira granulata, A. italica*, and *Fragilaria leptostauron* in association with fouling genera *Gamphonema* and *Cymbella.*

PALEOGEOGRAPHIC EVENTS AND THEIR CORRELATION

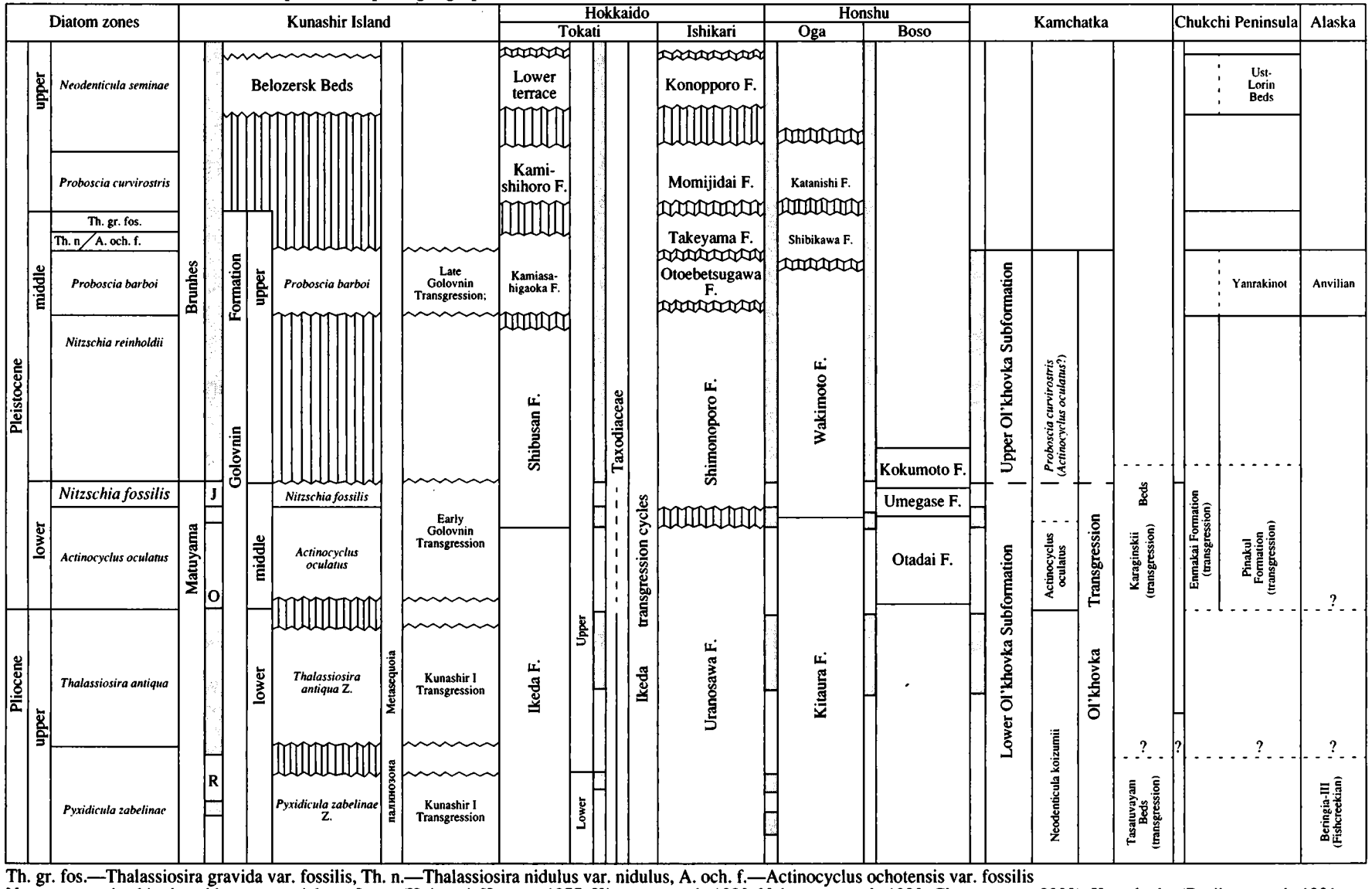
The Late Cenozoic history of Kunashir Island was characterized by unstable climate and eustatic sea-level fluctuations. Distinguished transgressive rhythms are regarded as the late Pliocene transgression with two the Kunashir I (2.12–2.0 Ma) and Kunashir II (1.8–1.6 ? Ma) phases separated by a regression, and as the early and late Golovnin transgressions of the early and middle Pleistocene. In fact, deposits of the Golovnin Formation accumulated only during the transgressions.

The interval of 2.2–1.9 Ma ago corresponded to a relative climatic warming and extensive transgression throughout the North Pacific, the sea level rise during which exceeded the present-day level by 50 m (Gladenkov, 1988; Basilyan *et al.*, 1991; Pushkar, 1998). According to Koizuimi (1994), annual temperatures of surface water in the NW Pacific at the end of the Pliocene exceeded the present-day values by 3.2°C that is close to calculations made by Barron (1992b). Using temperature coefficient Td (Koizumi and Kanaya, 1977), it is possible to assume that Δt was +2°C near the coast of Kunashir Island even at the early Olduvai time (Td_{old} = 19–20% and $Td_{pres.}$ = 17–18%). Climatic conditions at the end of the Pliocene were unstable, repeatedly fluctuating like those of the Tegelen time in Europe (Zubakov, 1990), of the Kutuyakh time in East Siberia (Laukhin, 1993), and of the Bering Land time (Sancetta and Silvestri, 1986; Bringham-Grette and Carter, 1992).

The diatom assemblage of the *Pyxidicula zabelinae* Zone with a high occurrence frequency of plankton species is a record of the Kunashir I transgression. This assemblage with subtropical elements is similar to present-day diatom phytocoenoses of the Kunashir region, but subarctic species present in it exclude a possibility to relate the assemblage origin to one of thermal maximums of the late Astian (Piacenzian). The assemblage most likely reflects temperate conditions of one of the climatic phases in Superclimathem 10 of the Tegelen Thermohyperclimathem (Zubakov, 1990). The subarctic front boundaries were similar to present-day ones. Diatoms from the *Neodenticula koizumii* Zone in upper parts of the Rybaki (?) (Section 3) and Parusnaya formations of the Iturup Island, which characterize the transgression cycle of the terminal Pliocene, may be similar in age to the diatom assemblage of the *Pyxidicula zabelinae* Zone (Vitukhin *et al.*, 1996, Oreshkina, 1996).

In the stratotype of the Maruyama Formation of Sakhalin, analogs of Assemblage 1 are upper beds of Member 5, which span the interval of reversed polarity and yield abundant planktonic diatoms *Proboscia barboi, Thalassiosira gravida* var. *fossilis, Stephanopyxis*

Correlation of Plio-Pleistocene deposits and paleogeographic events in the North Pacific



Th. gr. fos.—Thalassiosira gravida var. fossilis, Th. n.—Thalassiosira nidulus var. nidulus, A. och. f.—Actinocyclus ochotensis var. fossilis

Note: summarized in the table are materials on Japan (Koizumi, Kanaya, 1977; Kitagawa *et al.*, 1982; Nakagawa *et al.*, 1982; Cherepanova, 2002), Kamchatka (Basilyan *et al.*, 1991; Basilyan and Bylinskaya, 1997; Dolmatova and Nevretdinova, 1975; Oreshkina, 1993; Gladenkov, 1994), Alaska (Kaufman, Brigham-Grette, 1993; Pushkar *et al.*, 1999), and Chukchi Peninsula (Nevretdinova *et al.*, 1982; Ivanov, 1986; Polyakova, 1997; Pushkar', 1998).

horridus, and *Pyxidicula zabelinae* indicative of recession of the latest Pliocene transgressions in Sakhalin (Pushkar', 1983).

In the Karaginskii Island (table), Assemblage 1 is correlative with a warm-water diatom assemblage from the upper part of the *Neodenticula koizumii* Zone of the Tasatuvayam Beds, which is 2.2 ± 0.3 Ma old (Basilyan *et al.*, 1991; Oreshkina, 1993).

In the Chukchi Peninsula, traces of transgression corresponding in time to the second half of the *Neodenticula koizumii* Zone are probably recorded in the lower Enmakai and lower Pinakul strata, and in the upper part of the middle Rygytgyn Subformation (Polyakova, 1997; Pushkar, 1998). The diatom assemblage of the *Pyxidicula zabelinae* Zone characterizing the Bering Land III (Chukchi and Bering seas) and Fish Creek (Arctic Alaska) transgressions is distinguished in the Tuapaktushak Beds (Cape Barrow) of the Alaska Peninsula and in Bering III Beds of the Seward Peninsula (Pushkar' *et al.*, 1999).

In contrast to diatoms of the *Pyxidicula zabelinae* Zone, the assemblage of the *Thalassiosira antiqua* Zone marks the onset of warming and the new Kunashir II phase of the late Pliocene transgression after the abrupt global cooling. The interval of 1.9–2.0 Ma corresponds to a stratigraphic hiatus in deposits of the Golovnin Formation.

Traces of transgression under consideration have been found in lower members of the lower Ol'khovka Subformation of Kamchatka Peninsula, which reveal the normal polarity and accumulated in the Olduvai Time (Basilyan and Bylinskaya, 1997; Gladenkov, 1994). In the Karaginskii Peninsula, the marine diatom assemblage from the Karaginskii Beds near the Yunyunvayam River mouth is indicative of the same event (Volobueva *et al.*, 1992).

In the Chukchi Peninsula, synchronous events correspond to early phases of the Enmakai and Pinakul transgressions. Based on the *Proboscia barboi* and *Rhizosolenia matuyamai* diatom assemblages, they are dated back to the terminal Pliocene-Eopleistocene (Nevretdinova *et al.*, 1982; Polyakova, 1997).

Sediments of two diatom zones in the lower Golovnin Subformation (table) are correlated with middle strata of the Uranosawa and Ikeda formations of Hokkaido Island, which yield spore-and-pollen spectra with Taxodiaceae. The reference bed of the Chiyoda pumice flow separating the Uranosawa Formation into two parts are situated a bit above the Reunion Polarity Subzone, and the upper part of the formation (Osarushinai unit in recent terms) is as old as 1.46 ± 0.23 Ma according to results of fission-track method; the Plio-Pleistocene boundary is placed here at the Chiyoda level (Kitagawa *et al.*, 1982). Hence, the *Pyxidicula zabelinae* Zone corresponds to the Uranosawa Formation s. str., whereas the *Thalassiosira antiqua* Zone is an equivalent of the Osarushinai Formation. Well-rounded pumice from the lower part of the Khlebnikov Cliff section (Figs. 3, 4) is likely correlative with the Chiyoda pumice flow that is 2.0 ± 0.1 Ma old (Kitagawa *et al.*, 1982).

In the Neogene stratotype of Oga Peninsula, Honshu, the Pliocene-Pleistocene boundary is in the lower part of the Kitaura Formation, the boundary of which with the overlying Wakimoto Formation is placed at the level of the Jaramillo Subchron (Koizumi and Kanaya, 1977; Nakagawa *et al.*, 1982). Members of marine transgressive deposits in the lower part of the Kitaura Formation yield diatoms similar to the lower *Neodenticula koizumii* assemblage of the Golovnin Formation and correspond to the *Pyxidicula zabelinae* and *Thalassiosira antiqua* zones.

Several compelling analogs of the Early Golovnin transgression of the *Actinocyclus oculatus-Nitzschia fossilis* time are known. These are the Enmakai and Pinakul transgressions in Chukchi Peninsula (Polyakova, 1997), middle stages of the Ol'khovka (Karaginskii?) transgression in eastern Kamchatka (Basilyan and Bylinskaya, 1997; Gladenkov, 1994), the latest cycle of Ikeda transgression in Hokkaido, and the Otodai, Umegasi and Kokumoto transgressions in the Boso Peninsula of Honshu, all with proper paleomagnetic and biostratigraphic markers (*Proceeding...*, 1996). The *Actinocyclus oculatus* (the Otodai Formation and the lowest part of the Umegasi Formation), *Nitzschia fossilis* (the Umegasi Formation and the lower part of the Kokumoto Formation), and *Nitzschia reinholdii* zones (the upper part of the Kokumoto Formation) have been distinguished in the Kazusa Group based on presence of extinct species (Cherepanova *et al.*, 2001).

Diatom assemblages of the *Proboscia barboi* Zone from the upper Golovnin Subformation recorded two transgression phases coinciding in time with global oxygen isotope stages 11 (Sections 5800, 5838, 4695) and 9 (Sections 5837, 6095, 4595), during which the eustatic sea-level rise was as high as +20 m and +5 m respectively (Pushkar' *et al.*, 1998a).

In Hokkaido, upper transgressive members of the Golovnin Formation are identified with marine beds of the Otoebetsugawa Formation bearing warm-water molluscan fauna (Kitagawa *et al.*, 1987) and with the upper Wakimoto and lower Shibikawa strata of the Oga Peninsula, which yield similar diatom assemblages (Koizumi and Kanaya, 1977). In the Bering Sea sector, pronounced events of the Yarmutian Interglacial (OIS 11 and 9) are known as the Anvilian transgression (Cape Blossom Beds) in western Alaska, Yanrakinot transgression in the Chukchi Peninsula, and Wainwrightian transgression (Karmuk Beds) in Arctic Alaska (Ivanov, 1986; Polyakova, 1997; Kaufman and Brigham-Grette, 1993; Pushkar *et al.*, 1999). The diatom assemblage of the middle Pleistocene transgression was found in deposits of the upper Ol'khovka Subformation and the Lakhtak Formation of Kamchatka (Dolmatova and Nevretdinova, 1975; Gladenkov, 1994), but A.Yu. Gladenkov is of opinion that the nor-

mal polarity interval in the Upper Ol'khovka Subformation corresponds to the Jaramillo Polarity Subzone.

OIS 11 and 9 are well expressed in pelagic facies of NW Pacific (Jousé, 1962; Pushkar', 1998; Sancetta and Silvestri, 1986; Koizumi, 1994) and in various facies of marginal seas (Oreshkina, 1996; Cherepanova *et al.*, 1997; Pushkar', 1998).

The Pliocene-Pleistocene sequence of the Izmena Bay recorded three intense phases of the Golovnin Volcano eruption, which are represented by pumice pyroclastic flows of the Reunion (Khl-6, 7), Olduvai (Khl-5), and middle Pleistocene (Khl-II-4) ages (Fig. 2). Age analogs of these deposits are late Pliocene and middle Pleistocene (Kbl-II-1) pyroclastic flows composing the southern part of the Golovnin Cliff. It is likely that powerful directed outbursts of pyroclastic material were oriented southward and southeastward. Active volcanism in the southern Kunashir Island was synchronous with the activation of Hokkaido volcanoes in the Pliocene and middle-late Pleistocene (Okumura, 1988) and determined accumulation of thick tephroid and tuffite sequences on the shelf, the composition of which was inherited from reworked tephra.

Alluvial and lacustrine-alluvial sediments of the Belozersk Beds likely correspond to deposits of the Karaginskii (?) Interglacial of the late Pleistocene, though a lens of tuffaceous diatomite with the *Stephanodiscus-Aulacoseira* flora (the Belozersk Beds of Cape Puzanov) can be dated to the middle Pleistocene. It is apparent that there is a substantial stratigraphic and paleogeographic hiatus between deposits and events of the middle (Mindel-Riss) and late (middle Würm) Pleistocene.

CONCLUSIONS

Field and laboratory studies of the Golovnin Formation deposits in the southern Kunashir Island (the Kuriles) showed that the formation can be subdivided into three subformations, the origin of which was related to transgressions and periods of intense volcanism. Abundant and well-preserved diatoms represent the orthostratigraphic group suitable for detailed biostratigraphic subdivision of their host rocks. Deposits of the lower subformation divided into the *Pyxidicula zabelinae* and *Thalassiosira antiqua* zones formed during the Kunashir I (2.12–2.0 Ma) and Kunashir II (1.8–1.6 Ma) transgression phases of the Late Pliocene. Deposits of the middle subformation accumulated during the Early Golovnin transgression, indicative of which are diatom assemblages of the *Actinocyclus oculatus* and *Nitzschia fossilis* zones of the early Pleistocene. A long break in sedimentation was followed by a final transgression phase, when volcanogenic-sedimentary members of the Golovnin Formation were deposited. These members are attributed to the upper part of the *Proboscia barboi* Zone (0.43–0.28 Ma). Phases of the late Golovnin transgression coincide with OIS 11 and 9.

Based on paleoecological analysis of diatom assemblages, tephrostratigraphic and paleomagnetic markers, and on the event stratigraphy principles (periodicity of climatic changes, transgression cycles), deposits of the Golovnin Formation are correlated with Pliocene-Quaternary sequences of the Chukchi Peninsula, Alaska, Kamchatka, Sakhalin, Japan, and Pacific deep-sea standards. The trans-Pacific correlation showed that deposits of the West Pacific insular shelves can be regarded as a link between deposits of high and low latitudes, and between facies of marginal seas and pelagic zones of the ocean.

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Flora and Phytocoenoses of the Pre-Narev Pleistocene, Belarus

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Abstract—Based on carpological data, the composition and evolution of flora and phytocoenoses that existed in Belarus from the initial Pleistocene to the pre-Narev glaciation are analyzed and peculiarities of transformation of the late Pliocene flora into the glacial Pleistocene one are outlined. The analysis of the climatic affinity of flora and phytocoenoses and their evolution revealed the Vselyub Cryochron, El'nya Thermochron, Zhlobin Cryochron, and Rogachev Thermochron in the Early Pleistocene (Eopleistocene) succession of Belarus. The pre-Narev interval of the Middle Pleistocene is shown to comprise the Varangian Cryochron (glaciation) and Ruzhany Thermochron (interglacial). Substantial changes in the flora composition, particularly in the relict group, have been related to the Varangian glaciation and are recorded across the Early-Middle Pleistocene boundary.

Key words: Belarus, Early and Middle Pleistocene, paleocarpology, flora, phytocoenoses, thermochron, cryochron.

INTRODUCTION

In the specified Quaternary stratigraphic scheme with the lower boundary at about 1.8 Ma that is used in Belarus (Voznyachuk, 1981; Yakubovskaya and Nazarov, 1993; Velichkevich *et al.*, 1993, 1996; Yakubovskaya, 1998, *Geologiya Belarusi...*, 2001), the pre-Narev Pleistocene includes the Gomel Superhorizon representing the lower subseries of the Pleistocene (Eopleistocene) and the Brest Superhorizon at the base of its middle subseries (Lower Neopleistocene of Russia). The Narev glaciation is traditionally the initial one in the glacial Pleistocene. The stratigraphic scheme of the Gomel and Brest superhorizons is based on paleontological data, mostly on paleocarpological remains of prime importance.

Following L.N. Voznyachuk, we consider two main stratigraphic units of the pre-Narev Pleistocene of Belarus as superhorizons (Gomel and Brest) that include several horizons corresponding to cryo- and thermochrons (Table 1). Based on paleobotanical and paleoenthomological data, the Gomel Superhorizon is subdivided into the Vselyub, El'nya, Zhlobin, and Rogachev horizons and the Brest Superhorizon, into the Varangian and Ruzhany horizons.

The Quaternary section of Belarus shows two paleomagnetic markers. These are the Olduvai Subchron recorded in the clay member of the Upper Pleistocene near the village of Dvorets, the Rechitsy district of the Gomel oblast (Yakubovskaya, 1992), and the Matuyama-Brunhes boundary distinguished in the middle of the Brest Superhorizon that has been penetrated by Borehole 3 (=Borehole 13) near the village of Smolyarka, the Bereza district of the Brest oblast (San'ko and Moiseev, 1996).

The Lower-Middle Pleistocene boundary dated back to 0.8 Ma coincides with the base of the Brest Superhorizon. The Lower Pleistocene of Belarus is represented by continental analogues of the Apsheronian regional stage (Makhnach, 1977; Goretskii, 1977, 1980; Voznyachuk, 1985). In such a range, it corresponds to the Gurian regional stage in the Black Sea scale with its three horizons (Chepalyga, 1997) and Eopleistocene of the Russian Plain (Table 1).

The study of fossil flora from the pre-Narev Pleistocene sediments was started by P.I. Dorofeeev in 1966–1967 (the results are published only partly) and continued by many researchers (Velichkevich, 1973, 1982; Velichkevich *et al.*, 1993; Krutous, 1979, 1982, 1985, 1990; Yakubovskaya, 1978, 1984, 1985; Yakubovskaya and Rylova, 1992; Yakubovskaya and Nazarov, 1993). These studies established the floral composition in all horizons mentioned above.

Changes in the species composition of the pre-Narev Quaternary flora throughout its existence period (about 1 m.y.) reflect transformation of the late Pliocene flora into that of the glacial Pleistocene. This process in Belarus is unknown so far. Earlier works of Belarusian experts on paleocarpology showed that relict fossil taxa appear to be most suitable for study of flora evolution during the Late Cenozoic. Changes in their share in floral assemblages that form chronological succession accord with the Lyell-Ridow principle: the older is the flora, the higher is the proportion of extinct and alien genera and species in its composition. Until present, this principle serves as a most important tool for assessment of relative ages of fossil floras.

Table 1. Stratigraphy and correlation of the Lower-lower Middle Pleistocene of Belarus

Paleomagnetic scale, Ma (Berggren *et al.*, 1995)	$\delta^{18}O$ (Shackleton, 1997)	General scale Series, link		Belarus (Yakubovskaya, 1998) Superhorizon, horizon		Belarus (Voznyachuk, 1985) Superhorizon, horizon	Lithuania (Kondratene, 1996) Glacial, interglacial	Poland (Lindner *et al.*, 1998) Glacial, interglacial	The Netherlands (Zagwijn, 1989) Glacial, interglacial	Central Russia (Neopleistocene after Shik *et al.*, 2001; Eopleistocene after Iosifova, 1998)	
Brunhes	16	Pleistocene	Middle	Yasel'da –		Servech	Dzukiya –	San-1–	Glaciation B	Neopleistocene	Don –
	17			Korchevo +		Minsk	Bine +	Malopolie Interglacial – +	Westerhoven +		Il'inka + –
	18			Narev –		Narev	Nal'shya –		Glaciation A		Pokrov –
	19			Brest	Ruzhany +	Brest	Keméniai +		Waardenburg +		Petropavlovka – +
0.8	20				Varangian –		Kalviai –	Nida –	Dorst – Leerdam + Linge – Bavel s. str. +		
Matuyama	21–26		Lower	Gomel	?–+		Daumantai (sequence) – +	Podlasie Interglacial +		Eopleistocene	Port-Katon – + Ostrogozhsk +
	27–31				Rogachev +	El'nya					
	32–46				Zhlobin–			Narev –	Menapian –		Divnogor'e Formation – Novovoronezh –
1.8	47–49				El'nya +			Celestinov +	Waalian +		Nesmeyanovka +
	50–63				Vselyub –	Gomel		Otvock –	Eburonian –		Strelitsa-1– Kuteiniki –
2.6	64–103	Pliocene	Upper	Kolouchin	Dvorets +	Dvorets	Aniksciai (sequence) – +	Ponuzhica +	Tegelen +	Domashki +	
					Ol'khovka –			Ruzhce – +	Pre-Tegelen –		
Gauss	104				Kholmech +				Rever +		

Note: (+) interglacial thermochron; (–) cryochron.

The diagrams in Figure 1 demonstrate substantial differences in proportions of relict groups during the Gomel and Brest stages of flora development as a whole and during the separate thermo- and cryochrons. They demonstrate also sharp changes across the boundary between these stages and by transition to the oldest post-Narev flora of the Korchevo Interglacial. In total, 160 species are recorded in the pre-Narev flora. Among them, there are 60 (38%) species in common with the preceding late Pliocene flora of the Dvorets section. At the Gomel-Brest boundary, the number of relicts and particularly the share of extinct species sharply decreased: 36% of relicts amid 130 taxa in the Gomel flora (27% of them extinct) and only 30% of relicts (including 9% of extinct species) among 102 taxa in the Brest flora. Correspondingly, the proportion of recent species amid relicts increased. The relict forms also show the tendency of their abundance decrease during transformation of the Brest flora into the oldest post-Narev flora of the Middle Pleistocene Korchevo Interglacial: relicts of the Dvorets flora represent only 20% of 110 species registered in the last case, and the share of extinct species is only 4% (Yakubovskaya, 1978; Velichkevich, 1982, 1986).

The stepwise-changing rates of flora evolution by transition from the Early (Gomel stage) to Middle Pleistocene (Brest stage) and further to the interglacial flora of the Korchevo time are related to catastrophic events that suppressed relict elements. In our opinion, glaciations in the Northern Hemisphere could be among these events.

The past vegetation can be studied based on fossil floras. When reconstructing phytocoenoses for some intervals of the pre-Narev Pleistocene, we used paleocarpological and, partly, palynological materials along with data on geological and sedimentation settings in particular flora localities. Phytocoenoses and landscapes are reconstructed for the peaks of thermo- and cryochrons. The pre-Narev Pleistocene in Belarus was characterized by formation and repeated migration of vegetation zones from those covered by broad-leaved and mixed coniferous-broad-leaved forests with prevalent representatives of the present-day Belarusian flora to, probably, periglacial tundra-steppe with preponderant groupings of Arcto-Boreal Lycopodiaceae, some *Artemisia* and other forms, all alien to the present-day flora.

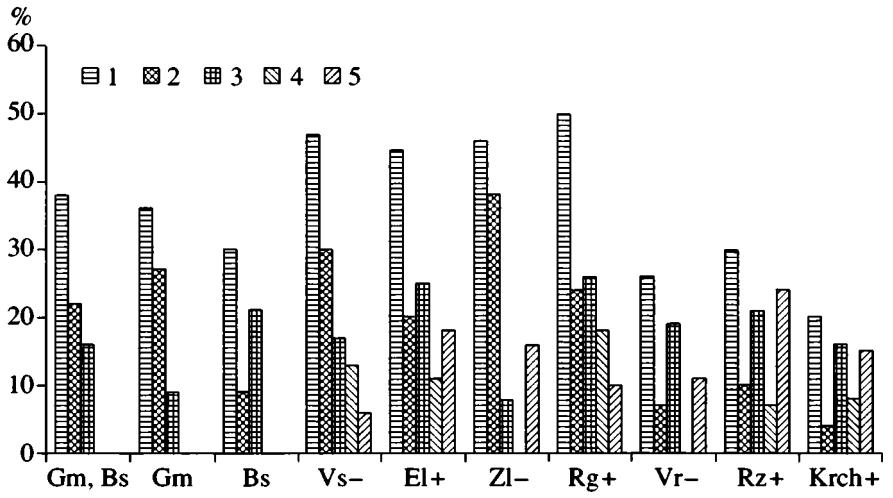

Fig. 1. Comparative abundance of relict and first appearing species in floras of the pre-Narev Pleistocene and oldest post-Narev (Korchevo) Interglacial of Belarus: (1) all relict taxa; (2) extinct relict taxa; (3) recent relict taxa; (4) LAD relicts; (5) FAD taxa; (Gm, Bs) combined Gomel and Brest stages; (Gm) Gomel stage; (Bs) Brest stage; (Vs–) Vselyub Cryochron; (El+) El'nya Thermochron; (Zl–) Zhlobin Cryochron; (Rg+) Rogachev Thermochron; (Vr–) Varangian Cryochron; (Rz+) Ruzhany Thermochron; (Krch+) Korchevo Interglacial. Abundance of species from groups of the last (LAD) and first appearance datums (FAD) are calculated for thermochrons (+) and cryochrons (–).

THE GOMEL DEVELOPMENT STAGE OF THE EARLY PLEISTOCENE FLORA AND VEGETATION

The Gomel stage corresponds to the entire Early Pleistocene. It comprises the Vselyub Cryochron, El'nya Thermochron, Zhlobin Cryochron, and Rogachev Thermochron. Floras of these subdivisions substantially differed in the species composition, particularly in composition and diversity of relict, first appearing, and extinct species. The units incompletely span the range of the Early Pleistocene, the terminal unstudied interval of which is indicated by question mark at the corresponding level of the stratigraphic scheme (Table 1).

The Lower Pleistocene subdivisions of Belarus are dated based on paleobotanical finds in the following reference sections: (1) Kholmech-3 section where oxbow clays 0.8–1.05 m thick are exposed in the buried terrace above the water edge of the Dnieper River in several localities between the villages of Kholmech and Dvorets in the Rechitsy district of the Gomel oblast (*Pliotsen Rechitskogo...*, 1987); (2) Borehole 141 that penetrated a peat layer in the depth interval of 156.0–157.8 m near the village of Mikelevshchina in the Mosty district of the Grodno oblast (Voznyachuk, 1978; Yakubovskaya, 1978); (3) Borehole 6 that recovered humic sandy loam (silt) and peaty gyttja in the depth interval of 29.8–37.8 m near the village of Khimy in the Rogachev district of the Gomel oblast (Velichkevich and Rylova, 1988); and (4) some other borehole sections (Fig. 2). The composition of the examined flora from all known localities is given in Table 2.

Vselyub Cryochron. This interval is the longest one in the Early Pleistocene of Belarus. It may approximately be correlated with Stages 50 to 63 of the oxygen-isotope curve (Shackleton, 1997). The cryochron is defined based on composition of the fossil flora and enthomofauna indicative of a sharp cooling (Yakubovskaya and Nazarov, 1993). This time corresponds to accumulation of the Vselyub Formation in the Niemen area (Goretskii, 1980) and its analogues in the Dnieper valley described there as the upper subformation of the Dvorets Formation (*Pliotsen Rechitskogo...*, 1987).

Paleobotanical materials characterizing the Vselyub Cryochron are known from sections of the Vselyub Formation (Makhnach and Rylova, 1977; Goretskii, 1980; Rylova, 1980; Yakubovskaya, 1984). Makhnach and Rylova (1977, p. 153) who analyzed the spore-pollen spectra from the section of Borehole 120 drilled near the village of Lozy in the Novogrudok district note "that the entire vegetation cover sharply changed at that time due to significant cooling caused by first advancing glaciation." The palynological data accord well with the seed flora that includes *Selaginella selaginoides, Potamogeton filiformis, P. vaginatus, Carex paucifloroides*, and other species indicative of a relatively severe climate.

The environmental conditions of the Vselyub time in southeastern Belarus are inferred from the detailed study of exposures in the left bank of the Dnieper River between the villages of Kholmech and Dvorets. In the Kholmech-3 locality (Zinova *et al.*, 1981; *Pliotsen Rechitskogo...*, 1987), the seed flora is characterized by a peculiar association of late Pliocene relict taxa, such as *Azolla interglacialis, Selaginella borysthenica, Pilularia pliocenica, Alisma plantago-pliocenica, Elatine pseudoalsinastrum*, and *Myriophyllum parvicarpum*,

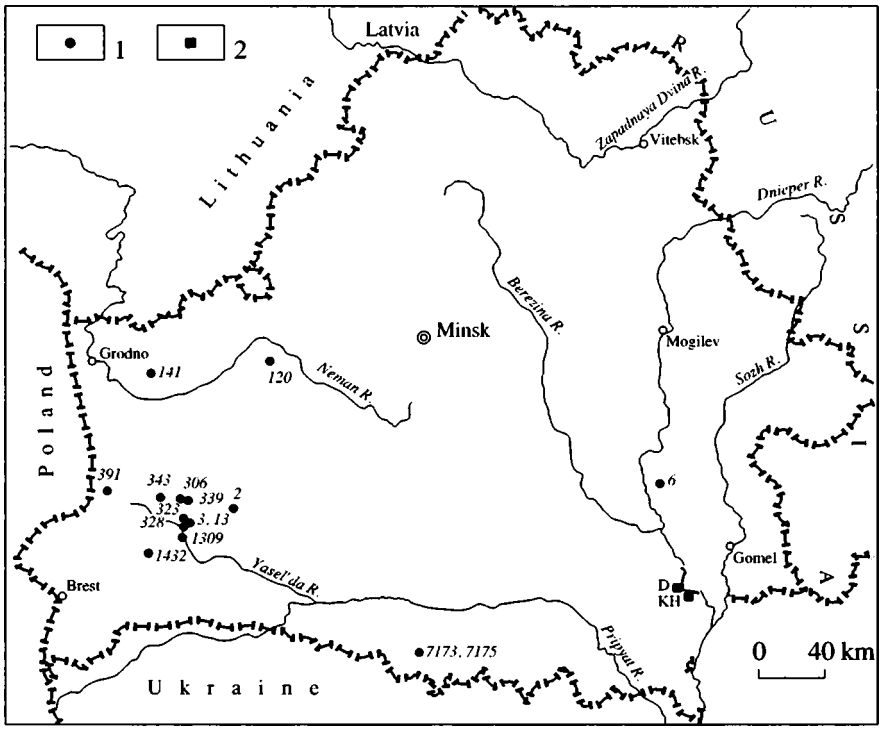

Fig. 2. Reference sections of Lower and pre-Narev Middle Pleistocene of Belarus: (1) boreholes; (2) exposures. Borehole localities by numbers: (*2*) Ivatsevichi in the Brest oblast; (*3, 13*) Smolyarka in the Bereza district of the Brest oblast; (*6*) Khimy in the Rogachev district of the Gomel oblast; (*120*) Lozy in the Novogrudok district of the Grodno region; (*141*) Mikelevshchina in the Mosty district of the Grodno region; (*306*) Iodchiki in the Pruzhany district of the Brest oblast; (*323, 328*) Bronnaya Gora in the Bereza district of the Brest oblast; (*339*) Kvasevichi in the Pruzhany district of the Brest oblast; (*343*) Verchitsy in the Pruzhany district of the Brest oblast; (*391*) Likhosel'tsy in the Pruzhany district of the Brest oblast; (*1309*) Strigin in the Bereza district of the Brest oblast; (*1432*) Postolovo in the Bereza district of the Brest oblast; (*7173, 7175*) Bukcha in the Lel'chitsy district of the Gomel oblast. Exposure localities: (D) Dvorets in the Rechitsy district of the Gomel oblast; (KH) Kholmech in the Rechitsy district of the Gomel oblast.

with species characteristic of glacial Pleistocene floras, such as *Selaginella selaginoides, S. helvetica, S. tetraedra, Potamogeton filiformis, P. vaginatus, Betula* cf. *humilis*, and others. According to these data, the Vselyub Cryochron only is characterized, in distinction from other pre-Narev Pleistocene cryochrons, by the complete set of the *Selaginella* species.

In the southwestern part of the region under consideration, the Vselyub stage is characterized by flora from the Borehole 1309 section (depth interval of 42.3–45.7 m) near the village of Strigin in the Bereza district of the Brest oblast. This flora is dominated by cold-resistant taxa accompanied by *Larix, Betula alba, B.* cf. *humilis*, and Arcto-Boreal herbaceous species (Velichkevich *et al.*, 1993).

In total, 55 taxa are registered in the seed flora of the Vselyub Cryochron (Fig. 1), 26 (47%) of which are the Pliocene relicts including 16 (30%) extinct and 10 (17%) recent taxa. Seven (13%) species disappeared within this interval: they became extinct or migrated from the study region precisely in the Vselyub time and never returned back. Only 3 (6%) species first appeared in the Vselyub sediments: these are *Selaginella tetraedra, Caltha palustris*, and *Cirsium palustre.*

These paleobotanical data enabled an adequate interpretation of the sedimentation setting and phytocoenoses in the ancient Dnieper River valley for the Vselyub Cryochron (Fig. 3). The river valley of that time inherited its Neogene morphology and was well developed, having a wide channel and abundant meanders. The river was characterized by a slow current and spacious flood. A shallow, as is evident from the thickness of filling sediments, although wide oxbow was incised into the Pliocene floodplain alluvium that composes a low terrace. The oxbow accumulated peripheral floodplain clayey sediments. The aqueous and coastal phytocoenoses were relatively uniform with some species being abundant. They consisted mainly of Potamogetonaceae, *Myriophyllum*, and Ranunculaceae species. The coastal areas of the oxbow were populated by of *Alisma, Carex, Scirpus*, Arundineae, *Sparganium*, and other typical Boreal macrophytes. *Selaginella* communities were characteristic of boggy areas in the marginal parts of terraces. The low watersheds composed of clayey sediments of the Miocene Antopol Formation and Paleogene marine sandy to silty sediments were covered by thin forests with preponderant birch (including dwarfish forms) accompanied by subordi-

Table 2. Composition of seed flora from the pre-Narev Pleistocene of Belarus

Species	Time					
	Vselyub	El'nya	Zhlobin	Rogachev	Varangian	Ruzhany
Azolla interglacialis P.Nikit.	—	—	—	—		—
A. pseudopinnata P.Nikit.		—				
Salvinia tuberculata P.Nikit.		- - - -				
S. glabra P.Nikit.		—				
S. aphtosa Wieliczk.				- - - -		
S. natans (L.) All. et var.		—	—	—	—	—
Selaginella borysthenica Dorof. et Wieliczk.	—		—	—		
S. reticulata Dorof. et Wieliczk.		—	—	—		
S. tetraedra Wieliczk.	—					
S. helvetica (L.) Spring.	—				—	
S. selaginoides (L.) Link et var.	—				—	
Pilularia pliocenica Dorof.	—					
Isofites lacustris L.		—			—	—
Pinus cf. *sylvestris* L., *Pinus* sp.	—	—		—		—
Larix sp.		—		—		
Picea sp.						—
Typha latifolia L.						—
Typha sp. sp.		—		—	—	—
Sparganium minimum Wallr.	- - - -			—		- - - -
S. emersum Rehm.		—				—
Potamogeton pectinatus L.		—				- - - -
P. filiformis Pers.	—			—	—	—
P. vaginatus Turcz.	—			—	—	
P. ex gr. *maackianus* A. Br.		—			—	—
P. praemaackianus Wieliczk.		—				
P. felixi Dorof.		- - - -				
P. ultimus Dorof.				- - - -		
P. crispus L.						—
P. palaeocompressus Dorof.				- - - -		
P. compressus L.		- - - -				—
P. acutifolius Link						—
P. praeacutifolius T.V. Jakub.	—					
P. trichoides Cham. et Schlecht.				—		
P. tenuifolius Raf.		- - - -				
P. ex gr. *coloratus* Hornem.	—					
P. pseudofriesii Dorof.				—		
P. friesii Rupr.						—
P. pseudorutilus Dorof.		—				
P. rutilus Wolfg.		—		- - - -	—	—
P. obtusifolius Mert. et Koch	- - - -	- - - -				—
P. obtusatus Dorof.				- - - -		
P. pusillus L.		- - - -		- - - -	- - - -	—
P. natans L.	—	—		—		—
P. stycovense Dorof.		- - - -				
P. lucens L.		—				
P. perfoliatus L.		- - - -			- - - -	
P. perforatus Wieliczk.		—	—	—		
P. praelongus Wulf.		—			—	—
P. densus L.					—	—
Potamogeton sp. sp.	—	—		—	—	—
Najas marina L.		—		—		—
N. major All.		- - - -				
Caulinia palaeotenuissima Dorof.		- - - -				—

Table 2. (Contd.)

Species	Time					
	Vselyub	El'nya	Zhlobin	Rogachev	Varangian	Ruzhany
C. interglacialis Wieliczk.		– – – –				
C. minor (All.) Coss. et Germ.		——				——
Alisma plantago-aquatica L.		——		——		——
A. plantago-pliocenica P. Nikit.	——					——
A. plantago-minima P. Nikit.	– – – –			——		——
Alismataceae gen.	——					——
Sagittaria sagitifolia L.				——		——
Stratiotes aloides L.						——
S. brevispermus Wieliczk.		– – – –				
S. intermedius (Hartz) Chandl.	——			– – – –	– – – –	——
Hydrocharis morsus-ranae L.				——		
Scheuchzeria palustris L.		– – – –				
Cyperus glomeratus L.						——
Scirpus sylvaticus L.	——					——
S. atroviroides Dorof.		——		——	——	——
S. lacustris L.		——		——var.		——
S. kreczetoviczii Wieliczk.				——	——	——
S. mucronatus L.						
S. triqueter L.						——
Eleocharis praemaximoviczii Dorof.	——	——		——		
E. ovata (Roth) Roem. et Schult.						——
E. palustris (L.) Roem et. Schult.	——			——	——	——
Carex blysmoides Dorof.			——	——		
C. paucifloroides Wieliczk.	——	——		——		——
C. pauciflora L.						——
C. cf. *vesicaria* L.					——	——
C. rostrata L.		——				——
C. rostrsta-pliocenica P. Nikit.	——			——		
Carex sp. sp.	——	——	——	——	——	——
Aracites inerglacialis Wieliczk.		——				
A. johnstrupii (Hartz) P.Nikit.		– – – –				
Lemna minor L.		——				——
L. trisulca L.				——		——
Betula alba L.		——		——		——
B. pendula Erdt.	——	——				
B. humilis Schrank	– – – –		——	——	– – – –	
Betula sp.sp.	——	——				——
Alnus sp. exot	——	——		——		
Urtica dioica L.		——		——		——
Rumex acetosella L.			——			
R. marschallianus Reichb.						——
R. maritimus L.	——			——		——
PolÛgonum aviculare L.						——
P. lapathifolium L.				– – – –		——
Chenopodium album L.					——	
Ch. rubrum L.	——	——				
Brasenia belorussica T.V. Jakub.		——				– – – –
B. cf. *holsatica* (Web.) Weberb.						——
B. borysthenica Wieliczk.		——var.				
Nymphaea alba L.						——
Nymphaea sp. exot		——		——		
Nuphar lutea (L.) Smith		——		——		
N. pumila (Timm.) DC.						——
Ceratophyllum demersum L.	– – – –	——		——		——

Table 2. (Contd.)

Species	Time					
	Vselyub	El'nya	Zhlobin	Rogachev	Varangian	Ruzhany
C. submersum L.						—
Batrachium aquatile (L.) Dum.	—	—	—	—	—	—
Ranunculus flammula L.					—	
R. lingua L.	- - - -			- - - -	—	—
R. polianthemus L.						—
R. reptans L.						—
R. acer L.	- - - -			- - - -		
R. pliocenicus Dorof.	- - - -					
R. sceleratoides P. Nikit. ex Dorof.	—	—				—
R. sceleratus L.	—	—		—		—
Caltha palustris L.	—			—		
Thalictrum lucidum L.				—		—
Rorippa palustris (L.) Bass.			—	—		
Rubus idaeus L.				- - - -		
Spirea sp.				—		
Comarum palustre L.	—	—			—	—
Potentilla anserina L.	—					—
P. pliocenica E.M. Reid	- - - -					
Potentilla sp. sp.			—	—		—
Filipendula ulmaria L.				—		
Crataegus sp.	—					
Frangula alnus Mill.		- - - -				
Euphorbia sp.				—		
Hypericum pleistocenicum Wieliczk.		- - - -				
Decodon goretskyi T.V. Jakub.		—		—		—
Elatine hydropiper L.						—
E. hydropiperoides Dorof. et Wieliczk.	—					
E. pseudoalsinastrum Dorof. et Wieliczk.	—					
Viola sp.	—					—
Trapa sp.						—
Myriophyllum spicatum L.		- - - -	—	—		—
M. praespicatum P.Nikit.	—	- - - -				
M parvicarpum Dorof.	—					—
M. ex gr. *microcarpum* Dorof.		—				
M. verticillatum L.				—	—	
Hippuris vulgaris L.	—	—		—	—	—
H. parvicarpa P. Nikit.	- - - -					
Oenanthe aquatica (L.) Poir.				—		—
Cicuta virosa L.				—		
Andromeda polifolia L.		—		—		—
Arctostaphylos uva-ursi L.	—					—
Naumburgia subthyrsiflora Nikit.						—
N. thyrsiflora (L.) Rchb.	- - - -	—		—		—
Menyanthes trifoliata L.		—		—	—	—
Lamiaceae gen.	—	—				
Lucopus europaeus L.				—		—
Mentha arvensis L.				- - - -		—
Stachys palustris L.						—
Sambucus cf. *nigra* L.						—
Valeriana sp.	—			—		
Bidens tripartita L.						—
Taraxacum vulgare L.						—
Cirsium palustre (L.) Scop.	—					

Note: Solid bars indicate presence of a species and dotted bars designate a close or cognate species; lines of relict taxa inherited from the late Pliocene are shaded.

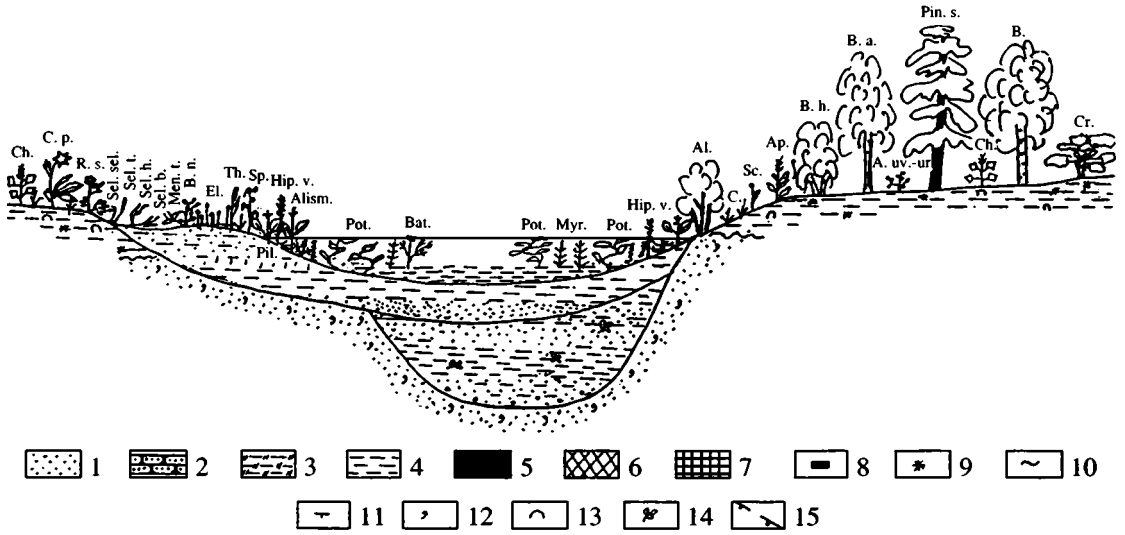

Fig. 3. Reconstruction of the water basin and phytocoenoses for the Vselyub Cryochron of the Early Pleistocene: (1) sand; (2) sandstone; (3) silt; (4) clay; (5) peat, coal; (6) sapropel, sapropelite; (7) chalk; (8) coal content; (9) variegated sediments; (10) humus content; (11) mica content; (12) glauconite; (13) kaolinite; (14) plant remains; (15) presumably active neotectonic fault zones. Abbreviations for plant taxa: (Alism.) *Alisma plantago-pliocenica, A. plantago-minima*, Alismataceae gen.; (Al.) *Alnus* sp.; (Ap.) Apiaceae gen.; (A. uv.-ur.) *Arctostaphylos uva-ursi*, (Bat.) *Batrachium* sp.; (B.) *Betula* sp.; (B. a.) *Betula alba*, (B.h.) *Betula humilis*, (B. n.) *Betula nana*, (C.) *Carex pauciflоroides, C. rostrata-pliocenica, C.* spp.; (Ch.) *Chenopodium rubrum*, (C. p.) *Comarum palustre*, (Cr.) *Crataegus* sp.; (El.) *Eleocharis* spp.; (Hip. v.) *Hippuris vulgaris*, (Men. t.) *Menyanthes trifoliata*, (Myr.) *Myriophyllum praespicatum, M. parvicarpum*, (Pil.) *Pilularia pliocenica*, (Pin. s.) *Pinus sylvestris*, (Pot.) *Potamogeton* (large); (R. s.) *Ranunculus sceleratoides, R. sceleratus*, (Sc.) *Scirpus sylvaticus*, (Sel. b.) *Selaginella borysthenica*, (Sel. h.) *Selaginella helvetica*, (Sel. sel.) *Selaginella selaginoides*, (Sel. t.) *Selaginella tetraedra*, (Sp.) *Sparganium minimum*, (Th.) *Typha* spp. (Pleistocene species).

nate pine, hawthorn, and alder (near water basins). The herbage cover between rare groups of trees and on terraces included the *Artemisia*-Chenopodiaceae and herbaceous communities. The landscape resembled a forest tundra and cold tundra-steppe. The environments of northern Belarus could be influenced by the Scandinavian glaciation of that time, which was probably responsible as well for loess accumulation and for occurrence of erratic detrital material in the Vselyub Formation.

The Vselyub Cryochron corresponds to loess-like loams of the Berezan time in Ukraine (Sirenko and Turlo, 1986), Eburonian of the Netherlands and the Otvock of Poland with a forest-free landscape. In the Oka-Don plain of Russia, its is correlative with the periglacial interval of the initial Eopleistocene that spans, according to Iosifova (1998), the Kuteiniki Superhorizon and Strelitsa-1 unit.

El'nya Thermochron. Under this term, Voznyachuk (1978) described the warm period, which he considered first as the late Pliocene in age. Later on, he attributed corresponding sediments to the Eopleistocene superhorizon (Voznyachuk, 1981). At present, the El'nya interval, as the warmest one in the Early Pleistocene, can be correlated with O-isotope Stages 47–49 and considered as representing a horizon in the Lower Pleistocene regional stratigraphic scheme of Belarus.

During the El'nya Thermochron, sedimentation took place in intensely overgrown and swamped river valleys and lakes to produce a local formation of organomineral and organogenic mud. Throughout the Belarus territory, the pre-Narev organogenic sediments were mostly deposited precisely in this period. The thickness of peaty and sapropelite sediments of that time is as great as 10 m.

Peat samples taken for the paleocarpological analysis from the Borehole 141 section near the village of Mikelevshchina in the Mosty district of the Grodno oblast yield the thermophilic flora (20 taxa) that includes *Picea, Betula, Salvinia natans, Azolla interglacialis, Aracites johnstrupii*, and abundant *Brasenia* seeds described as the Early Quaternary species *B. belorussica* with indication of another locality in the right bank of the Niemen River near the village of Nikolaevo in the Iv'ev district of the Grodno oblast (Yakubovskaya, 1978). The palynological study of the El'nya peat carried out by N.A. Makhnach suggests development of diverse forests with prevalent oak, pine, and Neogene relicts such as bog myrtle, wing nut, nut, and others.

The flora from Borehole 7173 drilled near the village of Bukcha in the Lel'chitsa district of the Gomel oblast (Yakubovskaya, 1989) is also referred to the El'nya period. In addition to relict and endemic forms (*Azolla pseudopinnata, Salvinia glabra, Selaginella reticulata, Hypericum* cf. *tertiaerum, Decodon*

goretskyi, and others), this seed flora includes a group of species recorded in the Korchevo and Aleksandriya interglacial floras of the Middle Pleistocene (*Stratiotes* cf. *brevispermus, Myriophyllum* cf. *praespicatum, Aracites interglacialis*). Yakubovskaya who discovered typical El'nya species *Decodon goretskyi* in Borehole 1432 (depth interval of 41.3–42.6 m) drilled near the village of Postolovo in the Bereza district also refers this flora to the thermochron under consideration (Velichkevich *et al.*, 1993).

The flora of the El'nya Thermochron is most peculiar amid other pre-Narev floras owing to presence of species that became extinct (*Azolla pseudopinnata, Salvinia glabra, Potamogeton praemaackianus, P. felixi*) and first appeared (*Brasenia belorussica, Aracites interglacialis, A. johnstrupii, Decodon goretskyi*) during this period. The flora composed of 71 species includes 32 Neogene relicts (45%), 15 (20%) of which are extinct and 17 (25%) represent recent species. Eight species (11% of the total assemblage) disappear and 13 forms (18%) first appear during the El'nya Thermochron (Fig. 1).

The thermophilic El'nya flora corresponds to the Early Pleistocene optimum in the Belarus territory.

The analysis of this flora, spore-pollen spectra from Borehole 141, and geological evidences characterizing sedimentation settings enabled reconstruction of the El'nya phytocoenoses in the Pripyat' River paleovalley that existed in this region since the late Oligocene (Fig. 4). The rich macrophyte flora with *Brasenia*, exotic Nymphaeaceae, *Caulinia, Stratiotes*, abundant Naiadaceae and Potamogetonaceae, aqueous ferns of the *Azolla, Salvinia*, and *Isotes* genera was accompanied by abundant algae and zooplankton responsible for deposition of sapropelic and organomineral sediments in oxbow lakes. The coastal herbaceous phytocoenoses were characterized by forms typical of the Middle Pleistocene interglacial periods, such as *Aracites, Scirpus atroviroides*, and others. Shrub thickets near river banks were composed of *Alnus, Salyx*, and *Myrica*. Dominant in mixed and broad-leaved forests were different oaks, pines, and Boreal amentiferous plants with subordinate share of spruce, larch, hornbeams, wing nut, and other Yuglandaceae. The available palynological data characterize only partly the forest phytocoenoses, which were undoubtedly more diverse in the southern part of the region in question.

The El'nya Thermochron of the Belarusian Early Pleistocene can be correlated with the Waalian Thermochron, the warmest period in the Netherlands, palynospectra of which point to two warming phases separated by a cooling event. Zubakov (1990) characterizes the Waalian Thermochron as the last period with ice-free summers corresponding in the Arctic to the last substantial degradation of the Antarctic ice sheet and sea-level rise by 20–25 m.

In Ukraine, this period is correlative with the Kryzhanovka Thermochron that was characterized by development of savannas and red soils under average annual temperatures of 10–11°C in the north (21–22 and 0–2°C summer and winter values, respectively) and precipitation of 600–800 mm/year (Sirenko and Turlo, 1986). These climatic parameters can be admitted to southern Belarus as well. For comparison, the present-day mean annual temperature in the Gomel oblast is 6.3°C (*Klimat Belarusi...*, 1996). In Russia, the El'nya time is represented by an upper interval of the Lower Eopleistocene corresponding, in opinion of Iosifova (1998), to the Nesmeyanovka Horizon of the Uspenka Superhorizon. In Poland, the unit in question is correlative with the Celestinov strata characterized by palynospectra with *Pinus, Betula, Picea, Alnus*, and *Quercus* (Baraniecka, 1991).

Zhlobin Cryochron The younger Lower Pleistocene sediments in Belarus are penetrated by Borehole 6 (depth interval of 29.8–37.8 m) near the village of Khimy in the Rogachev district of the Gomel oblast, where they are represented by humic lacustrine-boggy (oxbow) loam with an interbed of peaty-sandy gyttja and overlie clays of the Miocene Antopol Formation (Velichkevich and Rylova, 1988). Paleocarpological and palynological data show a transitional character of floras and phytocoenoses distinguishable between the Early and Middle Pleistocene. The Borehole 6 section reveals the Zhlobin and Rogachev intervals of the Gomel stage, which are named after nearby eponymous towns.

Sediments of the Zhlobin Cryochron are represented by humic sandy loam within the interval of 35.4 to 37.8 m. Analyzing composition of palynospectrum from this interval, Rylova noted "predominance of open partly swamped areas with phytocoenoses peculiar of late glacial tundra-steppe at that time" (Velichkevich and Rylova, 1988, p. 1015), and her interpretation seems to be valid. The paleocarpological characteristic of the Zhlobin Cryochron is based on a sample from the depth interval of 35.2–35.8 m, in which Velichkevich determined *Selaginella reticulata, S. borysthenica, Salvinia natans, Azolla interglacialis, Potamogeton perforatus, Carex blismoides*, and others (Table 2). Seeds characterize only 11 taxa, six of which (46%) are relict and represent five (38%) extinct and one (8%) recent species. Among taxa inherited from the Pliocene Dvorets flora, thermophilic forms are missing. The occurrence of *Betula middendorfi* and forms typical of areas with intense erosion imply development of forest-free landscapes that is consistent with palynological data. Based on paleobotanical evidences, the Zhlobin Cryochron may be characterized as a period of significant cooling and increasing aridity in southeastern Belarus.

In Ukraine, the corresponding period is marked by formation of the Il'ichevka loess horizon. Judging from pedological and palynological data, the climate of the Il'ichevka double cryochron was cold, more severe, as compared with that in the Berezan Cryochron: average

temperatures were 16 to 17°C in summer and –10 to –12°C in winter, while the annual precipitation was about 400 to 500 mm (Sirenko and Turlo, 1986). In the Black Sea history, this period corresponds to desalination stage related to ingression from the middle Apsheronian basin (Zubakov, 1990).

The Zhlobin Cryochron presumably correlative, in our opinion, with Stages 32–46 of the Pleistocene O-isotope scale corresponds most likely to a part of the Menapian glacial in Europe that is coeval with the Narev one in Poland (Lindner *et al.*, 1995) and with the Nebraska glacial (Zubakov, 1993) in North America. The cryochron also corresponds to Novovoronezh Horizon and Divnogor'e Formation of the upper Eopleistocene in Russia (Iosifova, 1998).

Rogachev Thermochron is defined in the interval of 29.8–35.4 of the Borehole 6 section, where it is represented by humic loam and peaty gyttja. These layers demonstrate a noticeable increase in abundance of arboreal pollen (up to 60–70%), although its content at some levels is lower, 15–16% only (Velichkevich and Rylova, 1988). These changes are justly considered as indicative of substantial warming and intermittent development of forests. During the entire Rogachev period, the forest communities competed with the meadow and steppe phytocoenoses distinctly dominated by *Artemisia* groupings.

Velichkevich was first to note a significant similarity of seed flora from this thermochron with the late Pliocene Dvorets flora of Belarus and with the Slavé-2 flora of Lithuania. The floras are characterized by the first appearance of *Scirpus kreczetoviczii*, the species whose maximum abundance is characteristic of the Middle Pleistocene Korchevo flora.

We tend to refer the seed flora from sediments penetrated at depth of 20 m by Borehole 7175 near the village of Bukcha to the Rogachev Thermochron as well (Yakubovskaya, 1989). Sapropelites from this borehole yield species *Selaginella reticulata, Pinus* sp., *Najas marina, Alisma plantago-aquatica, Carex* sp., *S. atroviroides, Nuphar lutea, Andromeda polifolia, Decodon goretskyi*, and *Menyanthes trifoliata*, most of which are present in the flora from Borehole 6 near the village of Khimy.

The flora of the Rogachev Thermochron consists of 71 taxa, 36 (50%) of which can be considered as the late Pliocene relicts representing 17 (24%) extinct and 19 (26%) recent species. Amid relict taxa, 13 species (18%) became extinct during this period, and 6 of 7 (10%) species first appeared at that time survived until present, whereas one species (*Scirpus kreczetoviczii*, which is unknown in the upper Pliocene sediments of Belarus), became extinct.

The moderately thermophilic affinity of the Rogachev aqueous-boggy flora and phytocoenoses is emphasized by occurrence of two *Salvinia* species in association with Potamogetonaceae, such as *Potamogeton perforatus, P. trichoides, P.* cf. *obtusus* and *P.* cf. *praecompressus*, and with *Stratiotes* cf. *intermedius, Hydrocharis morsus-ranae, Lemna trisulca, Nymphaea* sp. exot., and some other taxa.

According to Sirenko and Turlo (1986), the climate of coeval Shirokino Thermochron in Ukraine was warm, transitional to subtropical, and close to the Kryzhanovka one, but with slightly lower summer (20–21°C) and average annual temperatures (8–10°C) and with annual precipitation of 650–700 mm. Thus, it was comparable with the present-day climate of Belarus.

Paleocarpological data lead to the conclusion that the Rogachev flora is younger than the late Pliocene Dvorets flora, but close in age to the Slavé-2 flora (Velichkevich, 1973). The latter contains pollen of Pliocene relicts, such as *Tsuga* and *Pterocarya* (Kondratene, 1996) whose last occurrence in Europe is recorded in the Bavel Flora of the Netherlands (Zagwijn, 1989). We correlate the "interglacial" Rogachev interval of the Early Pleistocene in Belarus with the Bavel Flora range, although it would be more correct to correlate it with the Bavel s. str., i.e., with the first Bavel interglacial phase. The Rogachev Thermochron corresponds to the Ostrogozhsk Horizon of Russia in the scheme by Iosifova (1998) and, partially, to the Podlasie Interglacial of Poland (Lindner *et al.*, 1995).

THE BREST DEVELOPMENT STAGE OF THE MIDDLE PLEISTOCENE FLORA AND VEGETATION

In the geochronological scale of Belarus, the Middle Pleistocene begins with the Brest Superhorizon at the level of 0.8 Ma, i.e., its range is significantly widened. The Brest interval corresponds to a single climatic cycle consisting of the Varangian Glaciation (Cryochron) and Ruzhany Interglacial (Thermochron).

Varangian Cryochron (Glaciation). In our opinion, the Varangian sediments are correlative to Stage 20 in the O-isotope scale and their upper boundary corresponds to the paleomagnetic Matuyama-Brunhes reversal. The most comprehensive paleobotanical characteristic of the cryochron was obtained from the Borehole 13 section near the village of Smolyarka in the Bereza district, where these sediments were penetrated in the depth interval of 63.6 to 72.4 m (Yakubovskaya and Rylova, 1992; Yakubovskaya and Nazarov, 1993).

The alluvial sediments from the depth interval of 65.4–65.7 m yield the seed flora suggesting periglacial landscapes (*Selaginella selaginoides, S. helvetica, Potamogeton filiformis, P. vaginatus, Chenopodium album*, and others) and northern taiga biotopes (*Pinus* sp., *Menyanthes trifoliata, Myriophyllum verticillatum*, thorns of *Stratiotes* sp.). The flora also includes single Pliocene relicts (*Scirpus atroviroides*) and *Scirpus kreczetoviczii* that appeared in the Rogachev Thermochron. The spore-pollen spectra obtained by Rylova for several samples from the interval of 63.6 to 65.1 m in this section (Yakubovskaya and Rylova, 1992) do not

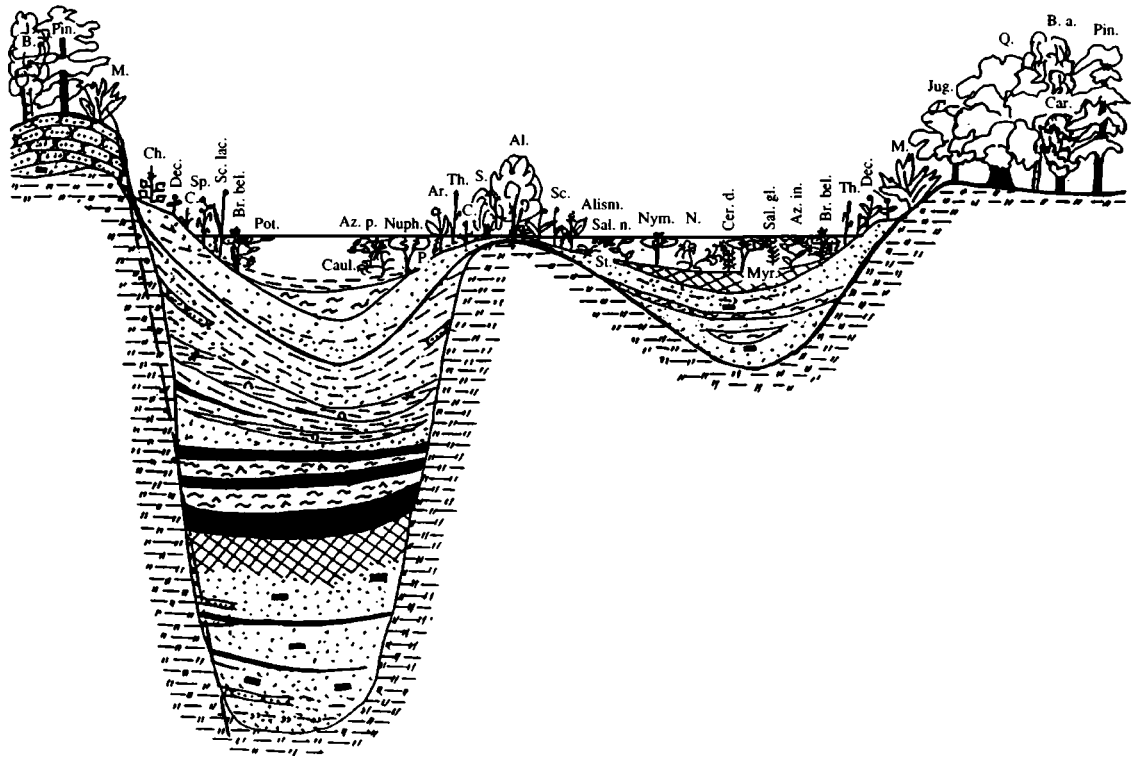

Fig. 4. Reconstruction of the water basin and phytocoenoses for the El'nya Thermochron of the Early Pleistocene (lithologic symbols as in Fig. 3).

Abbreviations for plant taxa: (Alism.) *Alisma plantago-aquatica*, (Al.) *Alnus* sp.; (Ar.) *Aracites interglacialis, A. johnstrupii*, (Az. in.) *Azolla interglacialis*, (Az. p.) *Azolla pseudopinnata*, (B.) *Betula* sp.; (B. a.) *Betula alba*, (Br. bel.) *Brasenia belorussica*, (C.) *Carex* spp.; (Car.) *Carpinus* sp.; (Caul.) *Caulinia palaeotenuissima, C. interglacialis, C. minor*, (Cer. d.) *Ceratophyllum demersum*, (Ch.) *Chenopodium rubrum*, (Dec.) *Decodon goretskyi*, (Jug.) Juglandaceae gen. (M.) *Myrica* sp.; (Myr.) *Myriophyllum praespicatum, M. spicatum*, (N.) *Najas marina, N. major*, (Nuph.) *Nuphar lutea*, (Nym.) *Nymphaea* sp. exot.; (Pin.) *Pinus* spp.; (Pot.) *Potamogeton* (large); (P.) *Potamogeton* (small); (Q.) *Quercus* sp.; (S.) *Salix* sp.; (Sal. g.) *Salvnia glabra*, (Sal. n.) *Salvinia natans*, (Sc. at.) *Scirpus atroviroides*, (Sp.) *Sparganium emersum*, (St.) *Stratiotes brevispermus*, (Th.) *Typha* spp. Lithological legend as in Fig. 3.

conflict with an assumption of severe climatic conditions at that time.

In 1965–1966, P.I. Dorofeev obtained fragmentary paleocarpological data characterizing the Varangian time. He found seed remains in sandy silt and sand penetrated by Borehole 343 near the village of Verchitsy in the Pruzhany district (depth of 83 m) and by Boreholes 323 (depth of 65 m) and 328 (depth interval of 74–75 m) drilled in the Bronnaya Gora coal occurrence of the Bereza district. These were remains of *Isoetes* sp., *Salvinia natans, Typha* sp., *Potamogeton filiformis, Carex* spp., *Batrachium* spp., *Hippuris vulgaris, Naumburgia subthyrsiflora*, and others.

In total, the Varangian flora contains 27 taxa, seven of which (26%) belong to Pliocene relicts, two extinct (7%) and five (19%) recent Belarussian species included. Since the Varangian time, the share of extinct taxa amid relicts sharply decreased. Like in the Zhlobin Cryochron, the Varangian flora lacks species disappearing at this level and includes three species (*Carex* cf. *vesicaria, Chenopodium album*, and *Ranunculus flammula*), which appeared here for the first time. Since that time, they are permanent component of interstadial floras and characteristic of non-optimal interglacial phases and of extraglacial landscapes during the glacial epochs.

The sedimentation setting and vegetable communities of the Varangian Cryochron (Fig. 5a) are reconstructed for southwestern Belarus based mainly on paleocarpological finds in sediments drilled by Borehole 13 near the village of Smolyarka in the Bereza district and on materials from neighboring boreholes. After the erosion phase of intense fluvial and, probably, glaciofluvial activity that left behind a member of sandy sediments, there was a phase of lacustrine transgression. Inasmuch as erosion affected the Neogene coal-bearing deposits in karst sinkholes of limestone land, it is possible to assume an active karstification in southern Belarus against the background of dominant ascending tectonic movements of the Varangian time. The subsequent lake transgression suggests tectonic subsidence of spacious areas in the Podlasie-Brest depression and northwestern part of the region under consideration.

The low coasts and islands of the Brest lake were occupied at that time by thin birch forests with preva-

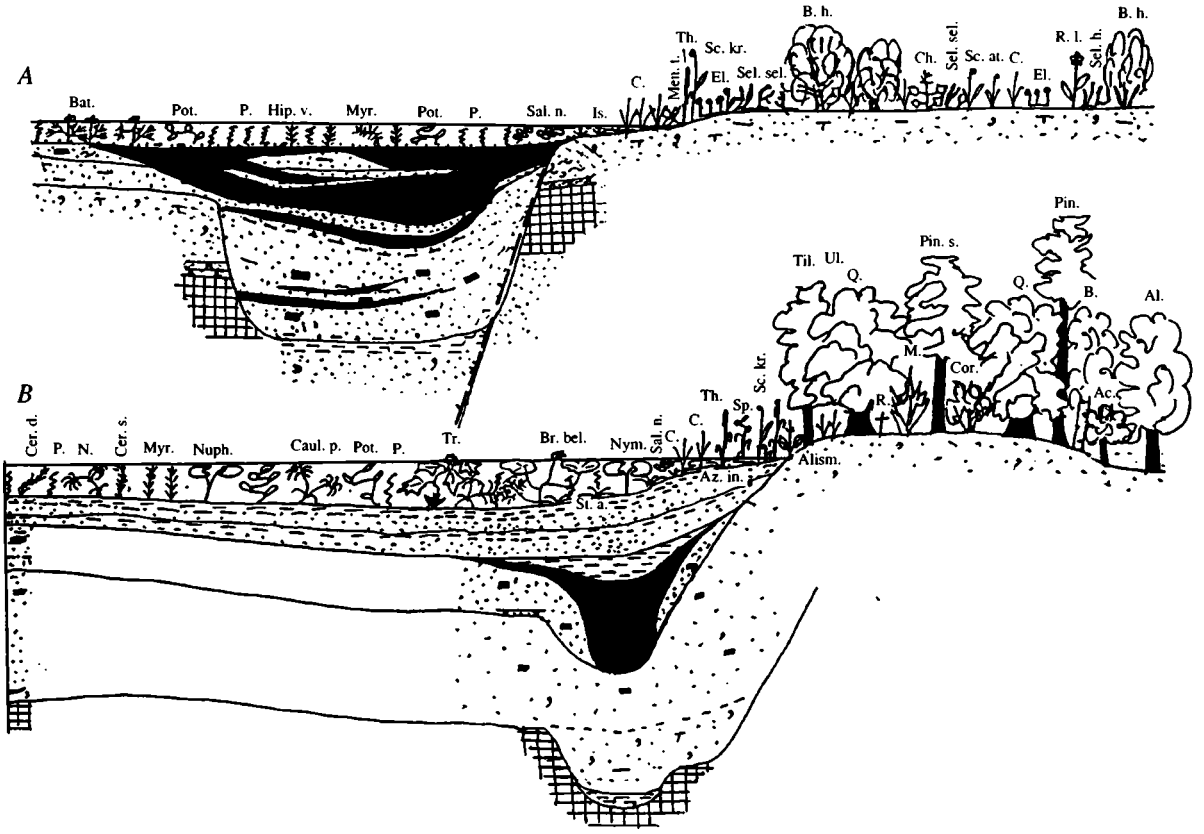

Fig. 5. Reconstruction of water basins and phytocoenoses for the Varangian Cryochron (*A*) and Ruzhany Thermochron (*B*), the Brest stage of the Middle Pleistocene (lithologic symbols as in Fig. 3). Abbreviations for plant taxa (A): (Bat.) *Batrachium* sp.; (B.h.) *Betula humilis*, (C.) *Carex* spp.; (Ch.) *Chenopodium album*, (El.) *Eleocharis* spp.; (Hip. v.) *Hippuris vulgaris*, (Is.) *Isoetes lacustris*, (Men. t.) *Menyanthes trifoliata*, (Myr.) *Myriophyllum verticillatum*, (P.) *Potamogeton* (small); (Pot.) *Potamogeton* (large); (R. l.) *Ranunculus lingua*, (Sal. n.) *Salvinia natans*, (Sc. at.) *Scirpus atroviroides*, (Sc. kr.) *Scirpus kreczetoviczii*, (Sel. h.) *Selaginella helvetica*, (Sel. sel.) *Selaginella selaginoides*, (Th.) *Typha* sp. Abbreviations for plant taxa (B): (Ac.) *Acer* sp.; (Alism.) *Alisma plantago-aquatica*, *Alisma plantago-pliocenica*, *A. plantago-minima*, Alismataceae gen.; (Al.) *Alnus* sp.; (Az. in.) *Azolla interglacialis*, (B.) *Betula* sp.; (Br. bel.) *Brasenia belorussica*, (C.) *Carex* spp.; (Caul.) *Caulinia palaeotenuissima*, (Cer. d.) *Ceratophyllum submersum*, (Cor.) *Corylus* sp.; (M.) *Myrica* sp.; (Myr.) *Myriophyllum spicatum*, *M. parvicarpum*, (N.) *Najas marina*, (Nuph.) *Nuphar pumila*, (Nym.) *Nymphaea alba*, (Pin.) *Pinus* spp.; (P. s.) *Pinus sylvestris*, (P.) *Potamogeton* (small); (Pot.) *Potamogeton* (large); (R.) *Rumex marschallianus*, *R. maritimus*, (Q.) *Quercus*. sp.; (Sal. n.) *Salvinia natans*, (Sc. kr.) *Scirpus kreczetoviczii*, (Sp.) *Sparganium emersum*, (St. a.) *Stratiotes aloides*, (Th.) *Typha latifolia*, *Typha* sp.; (Til.) *Tilia* spp.; (Tr.) *Trapa* spp.; (Ul.) *Ulmus*. sp.

lent shrub forms, *Betula nana* included. Open areas were populated by *Artemisia*-Chenopodiaceae herbaceous phytocoenoses. Swamped meadows included areas with *Selaginella selaginoides, S. helvetica*, and sedge groupings. Reeds were represented by *Scirpus sylvaticus, S. atroviroides*, and *S. kreczetoviczii*. Shoals in the water basin were occupied by pioneer phytocoenoses typical of still and slightly flowing waters and represented by *Potamogeton vaginatus, P. filiformis, Myriophyllum verticillatum, Hippuris vulgaris*, and others in association with abundant Ranunculaceae forms, *Elatine, Isoetes, Salvinia natans*, and *Azolla*. The landscape and phytocoenoses in southwestern Belarus of the Varangian time resembled the forest-tundra.

In the Vilnius and Miadininkai stratotype areas of Lithuania, the sediments correlative to the Varangian Glaciation are referred to the Kalviai Glaciation (Kondratene, 19996). They correspond probably to the lower part of the Petropavlovka Horizon in Russia, to sediments of the A or Dorst Glaciation in the Netherlands, and to the Nida Glaciation section of Poland, the upper boundary of which is marked by the Matuyama-Brunhes paleomagnetic reversal.

The Ruzhany Thermochron (Interglacial) is defined in the interval of 71–83 m of the Borehole 343 section drilled near the village of Verchitsy in the Pruzhany district of the Brest oblast located in the Ruzhany Pushcha (Yakubovskaya and Nazarov, 1993). The Ruzhany sediments are represented by clays and silts intercalated with sand layers. Dorofeev discovered the thermophilic seed flora with *Brasenia nehringii* (Web.) Szafer (*B. belorussica?*), *Cyperus glomeratus, Scirpus longispermus* Dorof. (*S. kreczetoviczii?*), *Stra-*

tiotes aloides, Lemna trisulca, and others at the depth of 78.0 m in the floodplain facies representing an upper part of the Brest sedimentary cycle. In the spore–pollen diagram compiled by A.F. Burlak for the interval of 77–83 m (four samples), this flora corresponds to the pollen assemblage (*Pinus-Quercus-Betula-Alnus*) from the upper sample.

Similar floras were described by Dorofeev in 1966–1967 from sections of Borehole 306 (depth level of 72.6 m) near the village of Iodchiki and Borehole 339 near the village of Kvasevichi (depth level of 82.5 m) in the Pruzhany district of the Brest oblast. Krutous (1990) recognized coeval seed floras in the Borehole 2 section near the town of Ivatsevichi (depth interval of 80.4–81.4 m) and the Borehole 391 section near the village of Likhosel'tsy in the Pruzhany district of the Brest oblast (depth interval of 108–111 m). These sections contain a peculiar flora characterizing vegetation of the Ruzhany time. It consists of 91 taxa, 28 (30%) of which are Pliocene relicts representing 9 (10%) extinct and 19 (20%) recent species. Some extinct late Pliocene taxa inherited from the Dvorets flora (*Azolla interglacialis, Scirpus atroviroides, Ranunculus sceleratoides*) continued to exist in Belarus during the Pleistocene up to the Aleksandriya interglacial period, whereas others (*Caulinia paleotenuissima, Alisma plantago-pliocenica, Stratiotes intermedius, Myriophyllum parvicarpum, Naumburgia subthyrsiflora*), which disappeared in the Ruzhany interval, were immediate ancestors of youngest Pleistocene and recent species. Among the last forms, the group of species that first appeared at this level is more diverse than in all preceding floras of the pre-Narev Pleistocene (22 species or 24%).

Arboreal plants are represented by local recent genera of coniferous *Pinus, Picea* and *Larix*, small-leaved *Betula*, and broad-leaved *Quercus*. Among aqueous-boggy herbaceous species, there is an outstanding group of thermophilic plants *Brasenia nehringii (B. belorussica?), Caulinia palaeotenuissima, Potamogeton acutifolius, Stratiotes aloides, Nymphaea alba*, and *Trapa* sp. The Ruzhany warming was characterized by flora typical of the interglacial periods of the Pleistocene. It was, probably, the warmest phase during the pre-Aleksandriya interval of the Middle Pleistocene.

The landscape of the Ruzhany Thermochron was covered by mixed coniferous–broad-leaved forests with *Quercus* prevalent amid broad-leaved forms, *Corylus* and *Acer* in the underwood, and *Alnus* and *Myrica* in riparian communities. The macroflora typical of the interglacial optimum (*Brasenia, Trapa natans, Najas, Caulinia, Stratiotes*, and large *Potamogeton* forms) populated thick mineral lacustrine grounds that overlapped the Miocene peat at that time. Similar thermophilic aqueous phytocoenoses developed only during the El'nya Thermochron of the Gomel stage.

In other areas of Belarus, sediments of the Ruzhany Thermochron are unknown so far. Their analogues are studied in southeastern Lithuania where they correspond, in our opinion, to deposits of the Vindziunai warming phase of the Keméniai Interglacial (Kondratene, 1996). The thermochron can be also correlated with one of two early interglacials of the Cromerian Group in the Netherlands and, probably, with an upper part of the Petropavlovka Horizon in Russia.

CONCLUSIONS

Peculiarities in evolution of the pre-Narev Pleistocene flora and phytocoenoses in Belarus enable recognition of two cryochrons and two thermochrons in the Early Pleistocene and of one cryochron and one thermochron in the initial Middle Pleistocene. This succession of climatic epochs spans only a part of the Quaternary. The complete range of some defined units remains unclear, and the history of the terminal Early Pleistocene in Belarus needs to be studied further to specify position of the Zhlobin Cryochron and Rogachev Thermochron, which seemingly represent analogues of the Menapian and Bavel s. str. Correlations with other regions and some other problems also wait their solution.

The landscape–climatic conditions that formed in Belarus during the Dvorets time of the terminal Pliocene (*Geologiya Belarusi...*, 2001) continued to exist in the Early Pleistocene as well. Mildest climatic conditions were characteristic of the El'nya Thermochron of the Early Pleistocene, when average annual temperatures exceeded the present-day values by 3–4°C and atmospheric precipitation was higher by 100 mm. Most severe conditions, similar to those characteristic of the glacial Pleistocene epochs, are reconstructed for the Vselyub time of the Early Pleistocene and for the Varangian period of the initial Middle Pleistocene, when forest-tundra and tundra-steppe landscapes could exist in the southern part of Belarus and periglacial tundra-like landscapes could occupy the northwestern part of the country.

The analysis of compositional evolution of relict species in the pre-Narev Pleistocene floras of Belarus elucidated criteria for the climatostratigraphic subdivision of this Quaternary interval and fossil flora age assessment. Par example, if the thermochron or cryochron flora contains 45–50% of relict taxa, it can be referred to Early Pleistocene Gomel stage of flora and vegetation development. When the share of relict forms in the flora decreases down to 26–30%, one can assume that these are floral remains of the initial Middle Pleistocene (Brest time). By dating the floras, the proportion of extinct relict taxa should be taken into consideration: it is 20–38% for the Gomel floras and 7–10% for the Brest floras. The floras of thermochrons and cryochrons differ also in the ratio between different relict groups and in proportion and composition of first appearing species. Floras of thermochrons always include all groups of relict taxa (extinct, recent, and last appearing species), whereas floras of cryochrons usually lack species, which disappeared during the relevant time inter-

val (Fig. 1). The pre-Narev floras of thermochrons are characterized by a higher proportion of the first appearing species, as compared with floras of cryochrons (10–24% versus 6–16% of the total amount of taxa). The substantial impoverishment of floras due to disappearance of species characteristic of the preceding epochs took place precisely during cryochrons.

Of extreme significance for a more precise age determination of fossil floras are the relict taxa of the pre-Narev Pleistocene and, in particular, the species of the last and first appearance datums (LAD and FAD groups), the latter representing characteristic taxa of the glacial Pleistocene. The following LAD relicts are known from the Early Pleistocene: *Salvinia glabra, S. aphtosa, Selaginella borysthenica, S. reticulata, Potamogeton praemaackianus, P. praeacutiformis, P. perforatus, Eleocharis praemaximoviczii, Carex blismoides, Elatine hydropiperoides, E. pseudoalsinastrum, Spirea* sp., *Myriophyllum praespicatum*, and *M. parvicarpum.* The diverse, although insufficiently studied FAD group includes the remarkable extinct forms, such as *Salvinia aphtosa, Scirpus kreczetoviczii, Aracites interglacialis, Brasenia belorussica, B. borysthenica, Decodon goretskyi*, and some others. Some of these species are also constituents of the LAD group of relict taxa.

In the Brest flora, the number of LAD taxa is significantly lower. These are *Alisma plantago-pliocenica, Caulinia palaeotenuissima, Naumburgia subthyrsiflora, Stratiotes intermedius*, and *Myriophyllum parvicarpum.* The FAD taxa are represented at this level by recent forms: *Potamogeton crispus, P. acutifolius, P. friesii, P. densus, Stratiotes aloides*, and others. According to composition and structure of the relict community, the Gomel flora is close to that of the Late Pleistocene and the Brest flora resembles floras of the glacial Pleistocene. This is as an indirect evidence of drastic climatic events during the Varangian Cryochron, when the ice sheet probably advanced into northwestern Belarus. In this connection, it would be more correct to refer only the Early Pleistocene to the preglacial period.

Reviewers N.Yu. Filippova and M.N. Alekseev

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The "Golden Spike" Method: A Problem of Geological and Physical Time Compatibility

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Abstract—The "golden spike" is a geological time mark at the axis of physical time. Such an interpretation is admissible only for particular sections, because the boundaries (lines) in the physical and geological time scales differ in essence from each other. In the physical time scale, they are virtual, inaccessible for direct observations, while the geological boundaries are real and classifiable by essence. Strategically, the GSSP (Global Stratotype Section and Point) method means a replacement of geological time (the succession of historical-geological processes) by the physical time standards in the International stratigraphic scale (ISS).

Key words: chronostratigraphy, geochronology, geochronometry, time, synchronicity, stratigraphic boundary.

The "golden spike" or the GSSP (Global Stratotype Section and Point) method, as it is usually termed, embodies, at first glance, a new idea in stratigraphy: the idea of time visualization by selecting a particular level (point) in a rock succession. In fact, it is sufficient to indicate particular, usually the reference sections to visualize the time recorded in them. Moreover, this approach is more suitable for that purpose, because several sections characterize the time better than a single one.

A peculiar feature of the "golden spike" method is the time visualization in a single section, i.e., in the lower boundary stratotype of the every ISS (International Stratigraphic Scale) unit, rather than in the geological record. This is substantiated by irreproachable logical consideration that correlation based on any stratigraphic features cannot guarantee a stratigraphic surface that is isochronous absolutely, and, consequently, the entire history of remote stratigraphic correlation can be viewed as a history of erroneous correlation. Any history is a process of changes, and a history of ideas (historical-geological here) is a history of "errors" in this sense.

An alternative proposed is the ideal project of objective fixing the points in particular sections to combine the physical and geological time instants. As a result, we receive an objective (at first glance) scale produced by coordinated activity of international working groups, i.e., a succession of reference points (GSSPs), which unite two kinds of time: geological and physical. In fact, these points are considered only as instants of physical time or as a basis for standardization: "A GSSP ... corresponds rather to a standard of measure in physics" (Remane *et al.*, 1997). This is understandable when variability of a geological time boundary (point in a section), which depends on our changeable historical-geological concepts, is taken into consideration.

To understand the essence of this procedure, one should consider the following issues:

(1) Is it possible to combine the physical and geological time in stratigraphic constructions? Can one speak at all about a geological time instant and isochronism of the given stratigraphic boundary?

(2) What is synchronism and accuracy in stratigraphy?

Both these problems are interrelated, and their solution depends on the fact whether we realize the essence difference between of geological and physical time or not.

TIME-DURATION AND TIME-PROCESS

It seems undoubted now that time is only one of constituting elements of the "substance-space-time" triad, but not an independent category. In the classical physics, time was considered first as something independent and in its quantitative aspect only, i.e., as a measure of the uniform movement (Galileo, Newton). Albert Einstein substantiated the time, space, and substance relationships first in his special (STR) and then in the general theory of relativity (GTR). The relationships are however rather formal because the substance in classical physics is only a gravitational mass. Plato, a philosopher of ancient Greece, was first who proposed to consider time as a measure of mechanical movement. Inasmuch as Plato did not admit the reality of material world, time was for him a mobile reflection of eternity, a movement from number to number. Aristotle, a materialist in his philosophy, created a more complete image of time: he related time to qualitative

changes of the moving object, but not only to its mechanical motion. Thus, two concepts of time were created in Ancient Greece: (1) time as an image of a uniform (monotonous) movement and (2) time as a process of qualitative changes.

In the first case, time is a succession of equal instants independent of any real processes (external time) and thus has been used as a basis for elaboration of progressively more accurate duration measurements (Galileo, Newton, Einstein). The second concept considering it as a succession of different intervals (stages) of the processes proper remained until recently a prerogative of philosophers and representatives of "inaccurate" (historical) sciences. In the second half of the 20th century, physicists discovered (rediscovered, according to I. Prigozhin) quantitative aspects of time when studying open, nonlinear, irreversible damping systems (processes), which include, undoubtedly, the historical-geological processes as well. It is paradoxical that stratigraphers, who accepted the chronostratigraphy concept simultaneously (the mid-20th century), staked on the accuracy, linearity, and objectivity, i. e. on the absolute time and isochronism.

The methodological discordance in development of sciences can probably be explained, despite their isolation, primarily by pure psychological causes: our everyday and internally nonuniform life proceeds against the background of monotonous independent time, the uniformity (accuracy) of which was permanently perfected by the science of the Recent time. The calendar and clocks became such a habitual attribute of our life that many stratigraphers consider a tool for measurement of time duration expressed in years or millions of years as identical to the time phenomenon itself. That is why Hedberg (1961), the main advocate of chronostratigraphy, was irritated by reasoning of opponents about a possibility of different understanding of the term "time."

Other details of methodological problems concerning the time category have been discussed earlier (Lazarev, 2002).

SIMULTANEITY AND ACCURACY IN STRATIGRAPHY

Isochronism is a key word in chronostratigraphy. Inasmuch as known correlation methods do not guarantee isochronism, the "golden spike" method is logically justified. In the present-day physics, simultaneity is relative: the site of an observer with chronometer corresponds to an interval (volume) between two light cones, not a line (surface), as the light velocity is limited (300000 km/s only).

There is no need to say that the propagation velocity of geological signals is several orders of magnitude lower, and precisely this indicates that the "isochronism" notion in stratigraphy has no and cannot have physical meaning. This is particularly true of the Newton's time that assumes the unlimited signal velocity in distinction from signals of geological time transmitted though the much more "viscous" medium.

However, this is not a main thing. An essential point is the principal difference between the physical time (t) expressed in years and the geological time recorded in rock successions as their stratigraphic features selected arbitrarily. The former is a faceless metrics of universal time useful for comparison of duration of different stratigraphic units, while the latter is a qualitative (topological) characteristic of intervals proper. The accuracy of the second time (geological or stratigraphic) is determined by possibility of using the smallest stratigraphic units (for instance stage, substage, or zone), not by a number of years.

It is psychologically difficult to consider synchronism in stratigraphy not thinking simultaneously about habitual units of the physical time scale (years). The word "scale" itself leads to a confusion: the determination of rocks' stratigraphic position is, strictly speaking, the classification, but not the measuring procedure. The accuracy of such procedure depends on the resolution degree of a stratigraphic scale (scale of geological time) and is determined by the possibility of using its smallest units in every particular case. Thus, the maximum accuracy in stratigraphy means the use of smallest units of the stratigraphic scale for determination of rock age, not the scale of physical time.

The same is true with respect to stratigraphic hiatuses. Observing distinct signs of a break in sedimentation, e.g., an uneven boundary of a bed with pebbles or a buried soil, in a section, where all zones of the standard scale are present, we cannot speak about a stratigraphic hiatus, although a gap in physical time is at hand, being equal to a unit of the corresponding local stratigraphic scale. Moreover, the bedding itself (boundaries between layers) is a result of some gaps in physical time. Meanwhile, the GSSP method as a tool for the ISS construction requires precisely the completeness of physical time in a section without reference to the quantitative assessment of this completeness. Whereas physicists understood relativity of their quantitative time estimates, stratigraphers who are dealing with qualitative (geological) aspects of time set hopes to the present day on the accuracy expressed in quantitative measurement units.

Indeed, it would be useful to have also data on the physical age of rocks expressed in years, but this would be the other time. I know three contemporary physicists (I.M. Dmitrievskii, U.V. Chashchikhin, and Yu.M. Stern) who arrived at the conclusion, independently from each other, that the Earth age is highly overestimated. The point is, I repeat, that the notion of "time" is dual in essence: it is, on the one hand, metrical, quantitative (expressed in years, seconds, and other units) and, on the other, topological, quantitative, classification-type (with qualitative measurement units such as an imago for insects, species for living organisms, zone for

stratigraphy). Contrary to metric models of physical time (for instance the Minkovskii model), there is no place for an observer with chronometer in the topological models (especially in the model of geological time visualized in the geochronological scale). The observer here is a classifier (a stratigrapher in geology) who uses a smallest appropriate stratigraphic "ruler" as a "chronometer." While physicist deals with features tied to points and lines, a stratigrapher immanently considers intrinsic, but not arbitrarily selected intervals of geological time. The seeming arbitrariness of the stratigraphic interval (the stratigraphic unit span) is epistemological (subjective), but not ontological in the essence: different stratigraphers differently assess the hierarchy of boundaries; moreover, even the uniformly accepted ranks of stratigraphic boundaries change with accumulation of additional data. Like everything in the material world, our knowledge is changeable.

Thus, the procedure of establishing the synchronism in stratigraphy is of the topological, i.e., of the classification, but not metrical, type (Krut', 1978), and synchronism implies the correspondence of rocks based on features, which are considered at the moment as stratigraphically most important (Meyen, 1989). In other words, synchronism in stratigraphy is not of a point type, as it always means a duration and a referral of rocks to the same historical unit based on their features in common (Krassilov, 1978). Thus, the notion of "synchronism" operates with volumes (intervals), but not with points or lines (surfaces). The latter should be considered therefore as classification (topological), but not as isochronous or diachronous lines, because their position is changeable depending on the state of knowledge. Regarding a stratigraphic boundary and boundary of physical time at any point selected in a particular section, we cannot speak about possible superposition of their continuations, because the essences of physical and geological time is different in principle. The boundary of physical time has nothing to do with a configuration of rocks, and, being continued (even if by extrapolation and interpolation), the stratigraphic boundary will turn into something that is observable in nature. Its position will be determined by a classification criterion, i.e., by features, which seem most important at the moment. It follows from the aforesaid that the "golden spike" method is methodologically unsound.

RELATIONSHIP BETWEEN GEOCHRONOLOGIC AND GEOCHRONOMETRIC SCALES

In connection with of the GSSP method of establishing points (instants) of physical time, two hypostases of time (physical and geological) are mixing in minds of many stratigraphers, as it is seen from terminology in use. It would be more logical to construct, from the very beginning of application of isotopic methods used to determine the age in years, an autonomous geochronometric scale tied to particular points and parallel to a traditional topological (classification) scale. In this case, it would be unnecessary to change the term "geochronometric scale" as it is defined in the Stratigraphic Code (1992, p. 22) for the term "scale of geological time" (*Dopolneniya...*, 2000, p. 83). This confusion resulted from mixing of two different aspects of time in a single "Scale of Geological Time" (ibid, Addendum 3, Tables 1–3). This title is not applicable to extreme right columns of the tables, where ages are given in million years. These columns represent solely a geochronometric scale. Precisely geochronometric units limited by instants of time must be marked by concrete points in particular sections-the GSSP analogues. Stratigraphic units of a true geological (historical-geological) scale, which are established based on correlation of the ultimate number of sections, do not need these points fixing instants of time.

One can combine, at least in his mind, the instants of physical time expressed in years with geological boundaries (changes in stratigraphic features) in any section. Using different methods (isotopic dates or stratigraphic features, which do not guarantee, in fact, isochronism) of the boundary lateral tracing from a fixed point, he will come in general to the resultant splitting of that boundary. Moreover, even the stratigraphic boundary will be paced at different levels depending on different preferable stratigraphic features. Nevertheless, regardless of the stratigraphic feature selected, one should not expect in principle its complete superposition with the boundary of physical time, because the latter cannot be observed in nature as a line (it can only be indicated as an observable point in a separate section). The "trick" of theorists in chronostratigraphy is in extrapolation of a real superposition of two time points (geological and physical) in a single section to an imaginable superposition of two time lines-the geological and physical boundaries. It is a due time to admit as a methodological utopia the dream that a reconstructed stratigraphic boundary in the ISS can approach the physically isochronous line encircling the globe. It is also remarkable that lexicon of stratigraphers is added by the following phrase: "this section bears a higher correlation potential." In fact, the correlation potential is immanence of stratigraphic features characterizing intervals below and above the selected point (boundary), but not of the section or point within it. Indication of particular sections (the more the better) is always necessary, but only as that of reference sections (analogues of experiments in exact sciences) for verification or falsification of stratigraphic constructions. Such a function of reference sections differs from the main function of the section with a "golden spike" (a carrier of a standard instant of physical time).

In connection with the last statement, A.Yu. Rozanov recommended me to consider the problem of such correlation methods, which seemingly imply isochronism in the geological sense (with an error insignificant for stratigraphy). Indeed, the rate of signals from different geological events can be highly variable. For

instance, it is clear that correlation based on ash interbeds can be considered as virtually proving isochronism. Considering however the ISS construction as a classification scheme, one should speak about historical-geological hierarchy of stratigraphic features, but not about their classification based on the relevant correlation accuracy. In addition, it is important that interregional correlation of ash interbeds (if the method of quantitative dating in years is meant) could hardly be possible, if stratigraphers had no topological (geochronological) basis (framework) created already for such a correlation. The same is true of instantaneity of geomagnetic polarity reversals.

When the commenced transformation of the ISS into a system of physical time standards is meant, I should agree with Rublev (1996) who proposed to supplement requirements to GSSP selection with the condition to date the boundary by isotopic methods and to consider this condition as the main one in selection of reference points as fixed instants of physical time. It does not matter that these points separate unequal time intervals; important is the exact localization of such standard points in rocks. Moreover, the accuracy of such scale is proportional to density of points at the time axis. It does not also matter whether the substantial overestimation of quantitative assessments will be shown in future or not. For stratigraphy, a relative remoteness of adjacent points (instants of time) is important, although the duration of intervals expressed in habitual years is also of interest.

The pathos of this paper is not directed against the use of isotopic dates of physical time, which always are extremely useful for stratigraphic positioning of fossil-free sequences or of those with poor stratigraphic features. But one thing is the positioning in the stratigraphic scale and quite another one is the scale construction itself using fixation of time points as standards of stratigraphic boundaries, which, first, cannot be isochronous in physical meaning and, second, are connected with the principally unstable temporary (in cognitive sense) hierarchy of geological time.

The quotation from the guide of the International Commission on Stratigraphy that is cited in the beginning of this article indicates that the essence of the GSSP method implies transformation of the scale of geological time into a scale of physical standards. For pragmatic stabilization of the ISS (as well as regional scales) without the geological essence change, it would be sufficient to reach agreement what stratigraphic features used to fix a particular boundary will be of first priority in the nearest future. Such a fixation of boundary has nothing to do with the GSSP method that represents a tool of constructing the exclusively geochronometric time scale. The latter is an autonomous scale, parallel to the geochronological one and not necessarily tied to stratigraphic boundaries, though it is necessarily tied to particular points in particular sections. Precisely the GSSP method promises us impossible: to lean upon a system of physical standards unchangeable in principle (a kind of a system of coordinates) instead of a subjective and unavoidably changeable knowledge of the geological time structure.[1] Such "standards" cannot be used in stratigraphy in a way as they are used in physics because the establishment of correspondence between any laterally remote points (correlation procedure) is a procedure of classification, namely the comparison of features located below and above some points in correlated section and ranking of these features. To a certain extent, this work implies eventually a procedure of historical-geological reconstruction (Meyen, 1989). In short, selected "standards" themselves cannot help to solution of the main stratigraphic problem, i.e., the correlation, because the latter is always connected with classification (topological) procedures, but not with quantitative measurements like in physics. We are doomed to deal with relative (inconclusive) stratigraphic concepts inconsistent with standardization. The paper by Popov (2002) published recently advocates an opposite ideology. It states, for instance, that correlation is a procedure of measurement and that the measurement represents a highest form of correlation. The author united physical time and time of organic world evolution into a single category and opposed it to the time connected with lithostratigraphy. The article by Popov is entitled *Stratigraphy as geochronometry*. Taking into account the aforesaid, I could propose the consonant title for my paper: *Stratigraphy is not geochronometry at all*.

CONCLUSION

First signs of anxiety concerning the GSSP method are seen in a recent paper *Should the "golden spike" glitter?* published by four authors abroad (Aubry *et al.*, 2000). The authors consider this method as deviation from the initially correct ideas of Hedberg. I have no doubts that the "golden spike" crowns logically the development of a main idea of chronostratigraphy that appeals to the time-clocks or time-duration. The "golden spike" hammered in the alive body of stratigraphy will remain a monument of a naive faith of stratigraphers in the philosophers' stone of stability and absolute. Every "golden spike" will glitter in the rock until the stratigraphic idea (level), which dominated at the moment of its fixation and did not anticipate a procedure of "spike hammering," preserves its allurement. As is noted with the reference to the guide of the International Commission on Stratigraphy, the "golden spike" is meant for elaboration of a system of physical time standards. As it is, it can remain eternally in the selected point, because it is not fated to fix the initial

[1] Time in the Einstein special theory of relativity depends on selection of a coordinate system (it is relative in this sense), i.e., the physical time itself implies a possibility of precise quantitative measurements and nothing else. This peculiarity of physical time (time-metrics) differs it from the time-process, i.e., from the geological time.

meaning (sense) of a boundary. It is of another strategic destination: to alienate from always-changeable pithiness and to move toward the objective and stable instant of time. On the one hand, it is "...a future anchor to which all subsequent correlations can be tied..." and, on the other, "...it is the only place where we actually know (by definition) that time and rock coincide in terms of our classification"; and further, the point is needed "...in order to define without doubt an instant of geological time..." (Cowie *et al.*, 1986). It is understandable that the adjective "geological" in this context becomes synonymous to the adjective "physical."

The absolute stability is pertinent only with regard to the geochronometric scale calibrated in years and closely tied to particular points in particular sections. The confinement of these points to boundaries in the historical-geological scale (par example, in the ISS) is desirable, but not necessary, because selection of reference points primarily depends on possibilities of the isotopic dating methods. Most important are the density of points at the physical time axis and their accurate correlation with particular sections. Such an approach provides an independence of a geochronometric scale from a geochronological (historical-geological) one, hierarchy of the latter is unstable in principle being determined by classification criteria. Correspondingly, boundaries of the ISS need not to be considered as isochronous or diachronous.

Our dream about the stable geological scale (scale-classification) turns gradually into prosaic destruction of the latter, and the GSSP method (with its stability ideal) justifies the nomenclature "brigandage" with respect to the Permian scale and its beginning with respect to the Carboniferous one.

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The Stratigraphic Code is a Collective Work (on Occasion of 25th Anniversary of the First Russian Stratigraphic Code)

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Key words: Stratigraphic code, stratigraphic classification, terminology, nomenclature.

"The USSR Stratigraphic Code," a first one published in this country, was approved by the Interdepartmental Stratigraphic Committee on May 10, 1976. The day later, I delivered its main principles at the 1st regional conference on Baltic stratigraphy held in Lithuanian Research Geological Prospecting Institute, Vilnius.

In May of 1977, the Stratigraphic Code subtitled "Provisional Regulations and Recommendations" was published in 15000 copies by the All-Russia Research Institute of Geology (VSEGEI). The further fate of the Code confirmed the well-known truth that there is nothing more permanent than something temporary.

The Stratigraphic Code presented basic regulations for using terms and names in the stratigraphic practice, and described procedure of establishing stratigraphic units. The code was appraised as one of the most important documents of the Russian geological survey that ensures to a possible extent "the uniform and stable employment of stratigraphic terms and names in different kinds of geological works in the USSR" (*Stratigraficheskii...*, 1977, p. 5). The code regulations were also kept in different geological investigations.

A scientific value of the code won a high appraisal either. When considering a problem of geochronological basis for geological investigations, B.S. Sokolov, Academician-Secretary of the Department of Geology, Geophysics, and Geochemistry evaluated the code in his annual report to the General meeting of the Russian Academy of Sciences on February 28, 1978. He said: *"The most significant event of the last year was probably the completion of many-year work of most qualified specialists of the Russian Academy of Sciences, Ministry of Geology, and high school who prepared 'The USSR Stratigraphic Code' published in many copies for a wide usage. The long-term scientific study and enormous practical experience ended in development of a single set of regulations and scientific criteria that determine the geochronological basis of the state geological mapping, mineral prospecting, and other geological works in the USSR. Strictly speaking, this work is even of a greater importance, because it offered a geochronological background for many global syntheses in geological cartography of different kind (tectonics, paleogeography, history of the Earth magnetic field, etc.). The new summarization of the Soviet stratigraphers differs essentially from that of the American stratigraphers, thus showing the importance of developing a single 'geochronological language' for geologists of the world."* (Sokolov, 1978, p. 7).

The code publication was a result of active long-lasted work of leading stratigraphers who analyzed a great deal of related publications in Russian and foreign languages. That is why A.I. Zhamoida, O.P. Kovalevskii, A.I. Moiseeva, and V.I. Yarkin are editors rather than authors of the Code.

Indeed, the code has grown from efforts of both living geologists and their predecessors. We should thank all those who have actively participated in this work. The preparation of the code and its subsequent edition represent a unique investigation, accompanied by heated discussions, which stimulated a great interest to theoretical and methodical problems of stratigraphy. The investigation grounded a progress in stratigraphic theory and practice.

The purposeful research of stratigraphic classification, terminology and nomenclature was initiated by setting up at the 1st International Geological Congress (IGC), Paris, 1878, a special commission aimed to unify stratigraphic and geochronological terminology. At the 2nd IGC session in Bologna, 1881, Russian participants proposed some terms for regional stratigraphic units, e.g., a suite (formation). The 3rd IGC session in Paris, 1900, approved the first taxonomical scale of stratigraphic and geochronological units and relevant terminology. The scale has become international.

Further results in this field of stratigraphy have been fully described in numerous monographs, textbooks, and various codes. Suggestions of three outstanding Soviet stratigraphers were basic in the preparation of the USSR Stratigraphic Code. Krishtofovich (1939, 1945) suggested two hierarchies of stratigraphic and geochronological units of global (general) and regional ranks. Stepanov (1946, 1951) pointed to distinction between the local and regional stratigraphic schemes. Keller (1950) proposed a hierarchy of local units (group, formation, member) and introduced the term

horizon for a local unit with a distinctive fossil content. Altogether, these researchers suggested a rather complete set of terms that were used in preliminary publications. The work was continued by the Stratigraphic Commission set up at VSEGEI in 1952 with L.S. Librovich as the leader. The result was the so-called "Green book" of 1954 ("Stratigraphic and Geochronological Units"), which immediately won a wide recognition. This book gave an impetus to the 1st All-Union Conference on General Problems of Stratigraphic Classification, which greatly advanced the stratigraphic thought and stimulated uniform approaches to all stratigraphic problems of regional and general significance (Sokolov, 1987, p. 7).

In 1954, the USSR government took a decision to set in action the state geological mapping at the scale 1 : 200000 and to issue many-volume monograph "Stratigraphy of the USSR." The Interdepartmental Stratigraphic Committee (ISC) headed by D.V. Nalivkin was organized in 1955 and issued a series of publications, among which there were books entitled "Stratigraphic Classification and Terminology" (*Stratigraficheskaya...*, 1956, 1960), "Compilation Instructions for General and Local Stratigraphic Correlation Schemes of the USSR" (*Instruktsiya...*, 1958), and "Study and Description of Stratotypes and Reference Stratigraphic Sections: Objectives and Regulations" (Librovich and Ovechkin, 1963). In 1959, V.V. Menner prepared a draft of "Regulations of Stratigraphic Nomenclature". Having been improved by a group of linguists, the draft was adopted at the ISC meeting in 1964 and included into "Stratigraphic Classification, Terminology and Nomenclature" (*Stratigraficheskaya...*, 1965) edited by Zhamoida, Menner, and Miklukho-Maklai.

In 1965, the ISC established the Commission on Stratigraphic Classification, Terminology and Nomenclature (CSCTN) with Zhamoida as the chairman. The commission was to prepare a stratigraphic code, basing on publications listed above. The first contribution was "A Review of Foreign Stratigraphic Codes" (Zhamoida *et al.*, 1969), which came out as the first volume of the ISC transactions.[1] The preliminary project of the USSR Stratigraphic Code was prepared by Yarkin in 1967. It was followed by two other code projects (*Proekt...*, 1970, 1974) and by a project of classification, terminology and nomenclature for the Quaternary system (Shantser *et al.*, 1973).

In addition to analysis of previous publications (including the foreign ones), the commission considered the stratigraphic correlation schemes and legends for geological map of some regions in the USSR and other countries. The well-known stratigraphers B.I. Bodylevskii, V.A. Zubakov, G.Ya. Krymgol'ts, L.S. Librovich, V.V. Menner, V.I. Yarkin, and others reported to CSCTN on principal stratigraphic problems in the context of the code compilation (*Proekt...*, 1970) at several meetings. Corresponding questionnaires on basic principles and regulations of the code were sent to various institutions.

What were the subjects of discussions? The necessity of a stratigraphic code that was doubted by few persons was soon out of debates, because the code was by no means intended to restrain development of the stratigraphic thought.

The controversial problems of stratigraphic classification and terminology have been discussed, based on opinions of leading stratigraphers, at the ISC meeting in 1967 (Zhamoida, 1969). Issues discussed at that meeting were as follows: relationships between stratigraphic and geochronological scales; stages in the organic world evolution and substantiation of stratigraphic boundaries; legitimacy of a single stratigraphic scale versus to combination of the general (global) and regional scales; and principles and criteria of establishing the general and regional (local) units.

The first project of the USSR Stratigraphic Code (*Proekt...*, 1970; 1000 copies) was commented by more than 70 experts (550 typewritten pages) and discussed by 50 participants of the ISC meeting on February 2, 1972 (Zhamoida *et al.*, 1973). The valuable remarks were made by Yu.N. Andreev, M.K. Apollonov, O.S. Vyalov, V.B. Goryanov, F.G. Gurari, V.E. Egoyan, G.Ya. Krymgol'ts, B.K. Likharev, N.P. Luppov, C.V. Meyen, D.V. Nalivkin, A.P. Rotai, A.M. Sadykov, T.N. Spizharskii, B.P. Sterlin, Yu.F. Chemekov, E.V. Shantser, Bureau of the Siberian Regional Stratigraphic Commission, and the Lithuanian Subcommission of the Baltic Regional Stratigraphic Commission. The views and opinions of Sokolov (1971) were very helpful. The majority of specialists highly appraised the project as a code base.

With respect to the purpose and content of the code there have been stated three alternative opinions: (1) the code must be dedicated only to the nomenclature regulations and relevant terms (S.V. Meyen and A.N. Oleinikov); (2) regulations for establishing units of any rank must be excluded from the code (O.S. Vyalov, G.Ya. Krymgol'ts and others); (3) the code must be transformed into a methodical guide. Nevertheless, the majority supported the project that included a table of stratigraphic classification, determination of principal notions and terms, and regulations for establishing and names of stratigraphic subdivisions.

The next main problem was definition of "stratigraphic units" (objects of stratigraphy). The majority defined them as primarily stratified bodies with distinctive properties elucidating a succession of their origin. L.L. Khalfin, T.N. Spizharskii, V.I. Krasnov, V.L. Egoyan, and some other geologists argued that any individualized bodies should be involved in the sphere of stratigraphy.

Owing to the previous discussions, the project proposals to reject "a single stratigraphic scale" (*Strati-*

[1] The English translation was published in Jerusalem (Zhamoida *et al.*, 1971) without authors' knowledge.

graficheskaya..., 1960), to regard stratigraphic units of different categories as independent, and to refer local units to principal (main) ones were approved without long debates and even termed as improvements to the former concept.

The problem of relationships between the general (correlation) and local units was one of central issues in the debate. More than 10 respondents (B.S. Sokolov, D.V. Nalivkin, G.Ya. Krymgol'ts, and geologists from the Research Institute of Arctic Geology and from the Institute of Geology, Geophysics and Mineralogy) attributed to both types a status of regional units. Other specialists, e.g., V.A. Vakhrameev and B.V. Poyarkov, argued that the category of correlation units should include the horizons and beds. O.S. Vyalov suggested "a regional stage" to be also included in this category as well.

As is known, the correlation units were ranked as regional ones in the second project of the code. Their correlation function was thought to be fitting the scale of sedimentation paleobasins or of paleobiogeographic areas of a comparable rank. At the same time, they were regarded as "bridging" the local and general units.

Discussions of that time about diachronism of boundaries separating the local units and about relationships these and auxiliary lithostratigraphic units are active even at present. The meeting approved a list of necessary addenda to the code. It was also decided that characterization of general units should be placed in the code before but not behind description of local units, as it was done to the project based on the French Code of 1962.

The main principles of the USSR Stratigraphic Code were published and reported to different organizations of the country and to the International Subcommission on Stratigraphic Classification at the 24th session of the International Geological Congress in Montreal (Zhamoida *et al.*, 1972). The report to the ISSC was published in the Circular 43 and found a response from foreign geologists.

In the second version of the project of the USSR Stratigraphic Code (*Proekt...*, 1974, 500 copies), definitions of terms were placed in the addendum and chapters dedicated to biostratigraphic and climatostratigraphic subdivisions were added. Over 50 comments to this version had been received, and then it was discussed at the CSCTN meetings in December of 1974 and in May of 1975. The version was approved as the basis for the future code (*Obsuzhdenie...*, 1977). It was also decided to introduce "a link" as a principal unit of the lowest rank, to retain regional units of the horizon and local zone ranks, to include the group into a category of local units, and to consider biostratigraphic subdivisions as independent units of particular substantiation. Because the majority of participants argued against description of climatostratigraphic units in the code, V.V. Menner was asked to head a working group that had to prepare an addendum devoted to units visualizing the periodic geological processes. The group has not been set up because of different reasons, but later Zubakov (1978) issued the book "Rhythmostratigraphic Subdivisions: A Suggested Addenda to the USSR Stratigraphic Code." Two of suggested units, namely, climatostratigraphic and magnetostratigraphic, were included into the second edition of the code (*Stratigraficheskii...*, 1992).

The final discussion on the project took place at the ISC meeting chaired by Menner (February 4, 1976), when, after report of Zhamoida, the meeting approved the third version for edition. The elected editorial board included A.I. Zhamoida (Chair), V.N. Vereshchagin, V.V. Menner, B.S. Sokolov, E.V. Shantser, and V.I. Yarkin. Together with O.P. Kovalevskii and A.I. Moiseeva, the board completed this collective work that lasted many years. The Code and its English translation were published jointly in 1979.

According to the ISC decision, the book "Stratigraphic Classification: Materials to the Problem" (*Stratigraficheskaya...*, 1980) was published. That book was a good opportunity for all stratigraphers to present personal views, including those, which contradicted the code.

Regulations of the USSR Stratigraphic Code were included into various methodical guides, such as instructions to geological surveying and mapping, the methodical textbook "Practical stratigraphy" (*Prakticheskaya...*, 1984), and high-school textbooks. They were in a permanent use by the Scientific Editorial Council of the USSR Ministry of Geology and its affiliates. It is likely that the code entered efficiently into practice of stratigraphic and geological investigations because many specialists from many regions and institutions of the country had been involved in preparation of the first code version. The code regulations became accustomed during the last decades, and geologists apprehended their necessity.

After the code publication, discussions on general problems of stratigraphy did not ceased but became even sharper. In the period of 1977–1988, there were issued more than 270 publications devoted to problems of stratigraphic classification and terminology, which were first outlined in the code. Many national stratigraphic codes and "International Stratigraphic Guide" were published abroad (Zhamoida *et al.*, 1996). In the USSR, the geological survey was oriented toward the geological mapping of the country at the scale 1 : 50000.[2] This necessitated the code perfection, and the work initiated by the ICSCTN in 1984 resulted in the second enlarged edition of the Code (*Stratigraficheskii...*, 1992) and in publication of "Addenda to Stratigraphic Code of Russia" (*Dopolneniya...*, 2000).

However, this is another story, as they say.

It is impossible to name here all the contributors to the code whose criticisms and even strong objections

[2] This work was cut down soon.

were very helpful. In addition to persons mentioned above, the detailed comments and proposals have been submitted by T.N. Alikhova, G.S. Biske, Yu.B. Gladenkov, V.P. Gorskii, A.A. Grigyalis, V.V. Drushchits, V.A. Zubakov, D.L. Kal'o, I.I. Krasnov, S.A. Muzylev, D.V. Nalivkin, V.P. Nekhoroshev, G.S. Porshnyakov, N.N. Predtechenskii, D.M. Rauser-Chernousova, G.P. Radchenko. L.I. Salop, Yu.V. Teslenko, A.V. Fursenko, I. Chlupač (Czechoslovakia), N.G. Chochia, and O.L. Einor.

After the code publication, I received many letters, occasionally critical but mostly encouraging. A first one was from Dmitrii Vasil'evich Nalivkin, the ISC Honored Chair, who was 88 years old then. His letter never published before is cited below.

Dear Aleksandr Ivanovich,

I wish to thank you for sending me a copy of the Code and much more for efforts to issue it.

I know that work was very complex, hard, time-consuming and unpleasant; it was fulfilled under your leadership and active participation, and regard this as a great achievement of you personally and of the VSEGEI as well.

Please, give my gratitude to your colleagues.

I favor the new term "link" (p. 20) for a subdivision of a zone. I think it will come into practice.

Now I am absorbed in describing activities of the first women-geologists and have already written more than 100 essays. Nevertheless, I am going to speak about "detailing of stratigraphic and paleontological units" at the VPO session, and the "link" is appropriate to be mentioned in my report.

Thank you once more

Yours D. Nalivkin

20-XII-77

My best wishes and happy New Year to you.

The authors and all the contributors to the USSR Stratigraphic Code were very pleased with the letter from the patriarch of Russian geology. Our many-year work showed that codes for geological disciplines must meet the following main requirements (Zhamoida *et al.*, 1996):

(1) As any code consists of regulations to be used in geological practice, it must be as brief as possible, intelligible, and well organized with a detailed table of contents. Any item of the code must be understandable without references to other items.

(2) Regulations to be included into a code must be those tested in practice and compatible with the world experience and national geological traditions; at the same time new tendencies must be taken into consideration; the obligatory regulation must be clearly distinguished from recommendations.

(3) A code is not a methodical guidebook or textbook, and its content cannot be restricted to a hierarchy of special units (subdivisions) and to regulations of their nomenclature. An essential part of a code must be aimed to define notions and terms, to characterize special units of different types (categories, kinds) and ranks, and to describe a procedure of their establishing and substantiation. Only basic notions and terms and a minimum of newly introduced ones must be used.

(4) The grammar rules of nomenclature, recommendations to practical study of type sections (exposures) and presentation of results, and any other information of practical significance may be given in addenda and notes.

(5) Theoretical problems of a certain discipline are not the objectives to be considered in a code, but it should be based on a certain concept and principles accepted by authors.

(6) The code regulations should be of a long-term stability, suitable for subsequent editions, because they are aimed to control practice of geological survey and mapping.

Whatever its explicitness would be, a code is unable to forestall all possible situations. For such a case, a good advice to use a common sense, when regulations or and recommendations fail to help, is given in both editions of International Stratigraphic Guide.

B.S. Sokolov (1978, p. 7) also suggested in his report cited at the beginning that "similar unification and standardization of notions, terms, and approaches to nomenclature are also necessary for other geological sciences (e.g., for tectonics, lithology, etc.)."

However, only the "Petrographic Code" (*Petrograficheskii...*, 1995), in which our methodic approach is taken into account, was compiled by the end of the last century on the initiative and under the leadership of N.P. Mikhailov, and a code of metallogeny is now under preparation in the VSEGEI.

There are different views on the necessity and possibility to work out a tectonic code. L.I. Krasnyi, who was long interested in problems of tectonic classification and terminology, came to the following conclusion: "After a preliminary discussion, it would be purposeful and adequate to formulate regulations and recommendations to tectonic systematics, terminology and nomenclature rather than to elaborate a tectonic code" (Krasnyi, 1994, p. 83). However, the items listed by Krasnyi are typical of all the codes. In stratigraphy, editions of this kind are not necessarily named codes. In Norway for instance, such an edition is defined as regulations. The International Subcommission on Stratigraphic Classification published two editions of International Stratigraphic Guide. The full name of the Russian code is "Stratigraphic Code of the USSR: Temporary Regulations and Recommendations". Thus, it seems quite possible to compile a tectonic code as well.

The successful experience of stratigraphers in the long-lasted work with stratigraphic codes should be

used, of course, by experts in metallogeny, tectonics, and other geological disciplines.

Reviewer Yu.B. Gladenkov

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Theory and Practice of Stratigraphy (on the Occasion of Russian Edition of the Abridged International Stratigraphic Guide)

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In 1952, the International Geological Congress held in Algeria adopted a resolution to prepare an international stratigraphic code. The term "code" was first introduced into stratigraphy by Australian geologists in 1950.

In the 1960s to 1970s, there was a boom in elaboration of national stratigraphic codes. The codes based on national traditions and experience in stratigraphic research have been issued in more than 25 countries (Zhamoida *et al.*, 1996). Some of the codes followed the European stratigraphic school of a single stratigraphy with the chronostratigraphic basis, while the others were oriented to the North American school declaring a certain independence of lithostratigraphy, biostratigraphy, chronostratigraphy etc.

At the same time, many outstanding scientists (Schindewolf, 1970; Erben, 1972; Laffitte *et al.*, 1972; Sokolov, 1971; Leonov, 1973; Menner, 1977; Menner *et al.*, 1977; Meyen, 1974; and others) initiated a spirited discussion on different approaches to classification of stratigraphic units.

The "International Stratigraphic Guide" edited by Prof. H. Hedberg was issued in 1976. This was a culmination point in the first period of the post-war activity of the International Subcommission on Stratigraphic Classification (ISSC) of the International Commission on Stratigraphy (ICS). The ICS approved the Guide for publication, but did not recommend to use it as a code because of its one-sided character and criticisms of some members of the Commission (Menner, 1977).

The guide gained a wide recognition in different, predominantly English-speaking countries and initiated a progress in the internationally acceptable approach to stratigraphic classification, nomenclature, and procedure. This facilitated the international communication between geologists and was favorable for mutual understanding.

The second edition of the guide edited by Prof. A. Salvador appeared in 1994. The document retained its general spirit, though with some important additions and corrections. For instance, the categories of magnetostratigraphic and unconformity-bounded units were introduced and biostratigraphic units received a more precise methodical interpretation. The described procedure of tracing the chronostratigraphic units (chronocorrelation) by using all possible methods was of extreme significance.

The first edition of the guide was translated into Russian in 1978 and widely distributed throughout the country (3800 copies), whereas the second edition is so far unknown to Russian geologists. It has not been translated into Russian, and only few English copies are available.

Journal "Episodes" (1999, vol. 22, no. 4) has recently published an abridged version of the guide second edition that was prepared by the ISSC and edited by M. Murphy and A. Salvador (*International...*, 1999). The editors believe that the abridged version is more obtainable (because of low price first of all) for stratigraphers and students in other branches of geology in different countries. The publication of the abridged version of the guide should be welcomed as the well-timed and useful for geologists of all countries.

The translation of the abridged version into Russian was supposed to be very helpful to our specialists. It was translated by Yu.B. Gladenkov, T.N. Golodnyuk, and O.A. Korchagin, all from the Laboratory of Phanerozoic Stratigraphy at the Geological Institute, Russian Academy of Sciences, and edited by Yu.B. Gladenkov and O.A. Korchagin. The document (*Mezhdunarodnyi...*, 2002) retains the structure of the original text and, what is very important, includes definitions of many terms commonly used abroad.

Prof. E.F. de Mulder, President of International Union of Geological Sciences, Prof. F. Gradstein, President of International Commission on Stratigraphy, and Prof. A. Salvador, editor of the International Stratigraphic Guide, encouraged the publication of the Russian version.

In view of criticisms of certain statements in the guide, the editors suggest to consider it as a recommended approach to stratigraphic classification, terminology and procedure, but not as a code. It may be supposed that in course of further international collabora-

Table 1. Categories and unit-terms in stratigraphic classification

Stratigraphic categories	Principal stratigraphic unit-terms	
Lithostratigraphic	Group Formation Member Bed, Flow	
Unconformity-bounded	Synthem	
Biostratigraphic	Biozones: Range-zones Interval-zones Lineage-zones Assemblage-zones Acme-zones Other kinds of biozones	
Magnetostratigraphic polarity	Polarity zone	
Other (informal) stratigraphic categories (mineralogic, stable isotopes, environmental, seismic)	Zone (with appropriate prefix)	
		Equivalent geochronologic units
Chronostratigraphic	Eonothem Erathem System Series Stage Substage (Chronozone)	Eon Era Period Epoch Age Subage (or age) (Chron)

tion this document can be used to develop an International Stratigraphic Code, which would be universally acceptable in spite of different national approaches to stratigraphic problems.

Certainly, at present the abridged version of the guide (like the guide itself) cannot substitute national stratigraphic codes, for example, the Russian Stratigraphic Code (*Stratigrafichskii...*, 1992) used successfully in stratigraphic study of the country. However, the acquaintance with this version allows us to evaluate different approaches to stratigraphy, especially to stratigraphic classification, on the one hand, and to understand some trends in development of stratigraphic science, on the other hand.

What are points of interest in the abridged guide? What are current approaches to stratigraphy and their strong sides? On the contrary, which statements have to be taken with caution or even rejected? Below is an attempt to elucidate these questions.

From the very beginning, I would like to indicate the merits of the version. They are as follows:

– *Economic wording* (a valuable feature for documents of this kind);

– *Logical structure* (identical to that of the full guide version);

– *Accessibility* (low cost for most geologists and students);

– *Pragmatic orientation* (importance for stratigraphic practice);

– *Clear definitions* of notions and terms (exclusion of ambiguous interpretations);

– Determination of *type biostratigraphic units* (understandable definitions and clear illustrations);

– Emphasis on *significance of regional stratigraphic schemes* (that has been often neglected in the last years);

– Formulated requirements for the *stratotypes* (of both the units and their boundaries);

– Elucidated relations between *different stratigraphic units*;

– Prime importance of *chronostratigraphy* stressed from the standpoint of stratigraphy main goal.

Leaving aside comments to self-understandable points mentioned above, I would like to list now the categories and units of stratigraphic classification, which are included in Table 1 of the Guide. These are *lithos-*

tratigraphic (group-formation-member-bed-flow), *unconformity-bounded* (synthem), *biostratigraphic* (biozones: range zones, interval-zones, lineage-zones, assemblage-zones, *acme*-zones and others), *magnetostratigraphic polarity* (polarity zone), *chronostratigraphic* (eonothem-erathem-system-series-stage-substage), and other (informal) subdivisions.

All these categories are considered as independent, although the chronostratigraphic units are stated to be the best ones for the worldwide use. Other units (lithostratigraphic, biostratigraphic and unconformity-bounded) are attributed to categories of regional significance. Magnetostratigraphic units are potentially of the worldwide importance, but they require data on other subdivisions in order to be dated.

In this classification, chronostratigraphic units are regarded as bodies of rocks formed during specified intervals of geologic time. In some cases, a category of chronohorizon, i.e., of a stratigraphic surface or interface that is isochronous everywhere, is admitted to be used. The classification also involves chronozones as formal chronostratigraphic units of unspecified rank, which are outside the general hierarchy.

Chronostratigraphic units of worldwide rank constitute a standard reference scale for dating all rocks and for definition of their place in world geological history. Boundaries of chronostratigraphic units are supposed to be synchronous (within the accuracy limits of existing methods of time correlation). The basic working unit of the lowest rank is the stage. As is stated, chronostratigraphic units and their boundaries can be correlated based on all possible evidences (lithological properties, paleontological content, isotopic age determinations, geomagnetic polarity reversals, paleoclimatic and eustatic changes, logging and seismic data, chemical and other properties of rocks).

In general, the above statements can be agreed with, but some of them need to be critically commented.

(1) The idea of independent litho-, bio-, magneto-, chrono-, and other stratigraphic subdivisions (*multiple stratigraphy*) seems to be incorrect. The equal-value status of above subdivisions is disclaimed by the authors themselves, when they note that chronostratigraphy uses data on all other stratigraphic units. In my opinion, the idea of *stratigraphy single in essence* seems to be more adequate, and we should make efforts to establish the veritable stratigraphic units ("chronostratigraphy" is a pleonastic term in relation to "stratigraphy," though acceptable in this case) based on combination of lithological, paleontological, paleomagnetic and other data.

(2) The definition of *chronostratigraphy* is somewhat uncertain. As is stated in the guide, chronostratigraphic units are bodies of rocks formed during certain intervals of geologic time, but what are these intervals proper remains unclear. The subdivision principles of ancient rock successions are not indicated. The guide specifies a procedure of tracing boundaries between the chronostratigraphic units but not indicates a basis they should be established on. It should be emphasized meanwhile that these units must be established *on the historical-geological basis only* as a visualization of certain evolutionary stages of the Earth and its blocks.

(3) The third problem is a disputable *classification* of stratigraphic units. Considered in the guide are several categories of units (lithological, magnetostratigraphic and others), which could be termed "special" like in the Russian Stratigraphic Code of 1992. According to classification in that code (Table 2), all the units are divided into the *principal* (they can be called "chronostratigraphic") and *special* ones. The former have an integrated substantiation and the latter are substantiated by particular data. According to their geographic range, the principal units are ranked as global (general), regional and local. The working special units correspond to lithological, biostratigraphic and other ones. Many of them are of practical use by the initial examination of sections. They supply data for substantiation of chronostratigraphic units of different rank.

(4) The next problem concerns *lithostratigraphy*. As is known, Schindewolf regarded it as protostratigraphy and French geologists as descriptive stratigraphy. To some degree, lithostratigraphic study can be considered as a preliminary stage of stratigraphy proper, because it is unable to yield the time information (clays or sandstones could be deposited either in the Silurian or in the Neogene). Owing to a frequent diachronism of lithological bodies, geological maps based on lithological properties are of a special, formational type, but not the structural ones. In distinction from them, the maps compiled on chronostratigraphic base show real structural patterns of different regions of the Earth.

(5) A very important problem of *regional stratigraphic scales* is worthy of a special attention. As correctly stated in the Guide, the route toward recognition of uniform global units lies through the local or regional stratigraphic scales. Moreover, the regional units will always be needed whether or not they can be correlated with the standard global units. Regional units are necessary because they are real records of regional history and serve as a basis for geological mapping and mineral prospecting. They are especially important in large territories situated far away from European stratotypes. Regretfully, the guide is *lacking consideration of hierarchy of regional and local units*, which, like global ones, should be ranked *on principles of chronostratigraphy* (and, therefore, are not identical to lithostratigraphic units whose boundaries are frequently diachronous). An example of such hierarchy is presented in the Russian Stratigraphic Code of 1992, and its practical usage deserves a serious attention. Regional units are the horizons (regional stages), lones (local zones), and beds with geographic name. Local units correspond to complexes, groups, formations, and members.

Table 2. Classification of stratigraphic and geochronological units (*Stratigrafichskii...*, 1992)

A. Principal stratigraphic units

General stratigraphic units	Geochronological units	Regional stratigraphic units	Local stratigraphic units
Acrothem	Acron	Horizon	Complex
Eonothem	Eon	(Subhorizon)	Group
Erathem	Era	Lone (local zone)	Formation
System	Period	Beds with geographic name	(Subformation)
Series	Epoch		Member
Stage (Substage)	Age		
Zone Division	Phase		
Link	Span		
Step	Thermochron		

B. Special stratigraphic units

Lithostratigraphic	Biostratigraphic	Climatostratigraphic	Magnetostratigraphic	Seismostratigraphic
Stratum Member Bed (layer) Marker horizon Organogenic massif Stratogene	Biozones of different kinds (Range-zone, Concurrent-zone, Phylozone, Interval-zone, Acme-zone, Oppel-zone, Assemblage-zone) Areal zones (provincial zone, local zone) Auxiliary units (beds with fauna or flora)	Climatolith Stadial Nasloi	Magnetozones (megazone, hyperzone, superzone, subzone, microzone)	Seismocomplex Seismohorizon

Possible continuation of Table 2B

Sequence-stratigraphic	Unconformity-bounded	Cyclostratigraphic	Ecostratigraphic	Others
Sequence	Synthem	Cycle (megacycle, supercycle, microcycle)	Ecozone	

(6) One more important fact should be mentioned. We frequently forget that chronostratigraphic units have *complex* many-sided characteristics. Every stratigraphic unit is characterized not by a single feature (for instance, paleontological or lithological one) but by a series of features, which define the unit *uniqueness.*

(7) It is to be noted that *chronozones*, which are not included into the hierarchy of chronostratigraphic units of the guide, have been long used in geological practice for elaboration of detailed stratigraphic scales. Sometimes chronozones are mapped and shown in geological maps. This raises a problem of introducing chronozone into the general hierarchy as a part of a stage. Although principles of choosing standard zones (among those based on different organisms) remains unclear for some systems and stages, this can be readily settled by mutual agreement. In the Russian Code, a zone is incorporated into the hierarchy of principal units as a part of a stage.

(8) In addition, the necessity of introducing zones into the general hierarchy is dictated by the fact that *detailed units (both zonal and infrazonal)* have now an increasing importance for geological practice. The latter are necessary for developing regional scales in order to subdivide borehole sequences and to compile large-scaled geological maps. Infrazonal units will gradually acquire an interregional and even subglobal application (Gladenkov, 1995).

(9) Finally, stratigraphy is often called a part of historical geology. It is aimed at subdivision of rocks in terms of their geological history and chronology that is necessary for development of the chronological scale and for dating of geological events. This is quite clear. Nevertheless, I would like to stress here that to subdivide rocks from the historical standpoint means to recognize the successive natural complexes, which recorded the successive *development stages* of individual regions or the whole Earth. In a broad sense, stratigraphy turns into the science of successively changing ecosystems.

The main goal of stratigraphic investigations is to elaborate stratigraphic scales and schemes of local, regional and global significance.

In addition, in the sphere of stratigraphy there are significant problems of general geology, primarily the recognition of natural periods in geological history of the Earth (and biosphere) and geological mapping. The former makes stratigraphy an independent part of geology. The latter illustrates a practical significance of stratigraphy.

It is not surprising therefore that, along with the traditional problems (stratigraphic classification, stratigraphic scales, and so on), stratigraphy must study *the history of biosphere* (paleoecosystems) and manifestations of geological processes (that is why present-day stratigraphy may be termed the "biospheric" one). It is not incidental that modern stratigraphy went outside the methodological-technological problems (establishment of stratigraphic units and their boundaries, etc.) and, analyzing "the past biospheres," it studies the Weltanschauung problems discovering *natural stages of the Earth geological evolution*, trends, and succession of different-rank geological events. In this respect, stratigraphy is not an applied science only. It is a science of a general methodological significance for humanity, and its role in education of young generation cannot be overestimated. Being deprived of stratigraphy with its geohistorical concept, geology would be a descriptive science to a considerable extent.

Since the forthcoming objective is to perfect the International Stratigraphic Guide and even to work out an *International Stratigraphic Code*, a discussion of the mentioned problems by geological community seems quite necessary.

Reviewer I.A. Basov

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